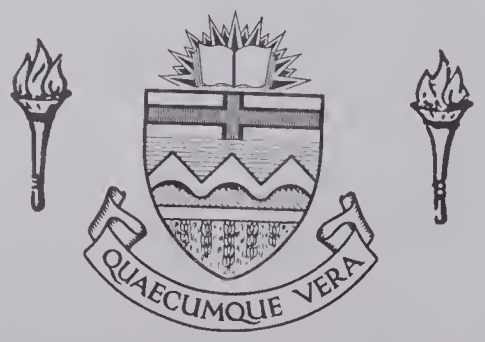
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ASPECTS OF CHANNEL AND BASIN MORPHOLOGY IN THE STEVEVILLE BADLANDS, ALBERTA

by

© PATRICIA HAZEL FAULKNER

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES

IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE

OF MASTER OF SCIENCE

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies for acceptance, a thesis entitled "Aspects of channel and basin morphology in the Steveville badlands, Alberta", submitted by Patricia Hazel Faulkner in partial fulfillment of the requirements for the degree of Master of Science.

ABSTRACT

Extensive badlands have developed peripheral to the Red Deer River in southern Alberta. Exposed on the retreating valley walls, bentonitic clays are being eroded by convectional storms in the spring and summer favouring the rapid development of gullies which extend from the river into the adjacent upper prairie.

Gully evolution in the Steveville vicinity can be illustrated sequentially by examining the hypsometric curves and stream profiles of ten tributary gullies of increasing size. The series of stream profiles and basin forms represent a model illustrating changes that occur as the basins enlarge and extend laterally in the upper prairie where no headwater competition occurs. In order to assess whether the planimetric characteristics of the basins similarly change, the drainage systems can be ordered and consistency in the areal division of the networks examined from the relevant Hortonian regressions.

The characteristics of the ten channels as they emerge from their watersheds were mapped in the field, and cross-sections measured on each channel segment. The magnitude of bankfull discharge can be assessed, and this value and other channel characteristics correlated with drainage net variables. Exponents of plotted regressions can be compared with those obtained in other areas, deviations being explicable within the presented model. The mode of gully extension in the Steveville area, and changes that take place in both hydraulic and morphometric relationships as this extension proceeds are thus elucidated.

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LIST OF SYMBOLS

A,B,C,D,E etc.,	Major structural zones, see Fig. 3
a,c,k,p,b,f,m,	Constants and exponents used in hydraulic geometry theory
$\bar{A}_o$	Average area drained by all streams order 0 in a basin
A_c	Cross-sectional area of a channel
$\bar{A}_c$	Average cross-sectional area of a channel
A_d	Drainage area, general term
A_s	Planimetric area of a basin order s drained by main channel length L_s
d	Channel depth. General term, although usually referring to depth at bankfull
D_d	Drainage density, = $\Sigma L/A_s$
GR	Gradient of main channel, = L_s/H
H	Total elevation of a basin
$\bar{L}_o$	Average length of all streams order 0
L_o	Total length of all streams order 0
L_B	Maximum basin length from mouth
ΣL	Total length of all stream segments in a basin
L_d	Length of main channel, general term
L_s	Total length of main channel draining basin order s
n	Manning's roughness coefficient
N_o	Total number of all streams order 0
O	Basin order (Strahler)
P	Wetted perimeter
Q	Discharge

Q_b	Bankfull discharge
Q_{bc}	Calculated bankfull discharge
$Q_{2.3}$	Mean annual flood (equalled or exceeded on the average once every 2.3 years)
r_A	Area ratio, $= \overline{A}_o/\overline{A}_{o-1}$
r_b	Bifurcation ratio, $= N_o/N_{o-1}$
r_L	Length ratio, $= \overline{L}_o/\overline{L}_{o-1}$
r_e	Elongation ratio, $= \frac{\text{diameter of circle same area as basin}}{L_B}$
R	Hydraulic radius
$ss_{1,2,3,}$	Indurated sandstone bands, 1,2,3, (See Fig. 3)
s	Highest order stream in a basin. Basin is allotted this order
S	Slope of energy grade line through a cross-section. Equivalent to channel slope through the cross-section. Refers to bankfull
V	Velocity
V_b	Bankfull velocity
V_{bc}	Calculated bankfull velocity
w	Channel width. General term, but usually refers to bankfull
W_1	Suspended sediment load

INTRODUCTION

Badlands provide geomorphologists with unique environments in which to study landform development. Rapid erosion on a surface usually unvegetated and uncomplicated by human interference makes it possible to isolate relationships between landforms and their controlling processes with greater ease than is possible under most conditions. In areas where climate and geology are relatively uniform, these relationships can be extrapolated both areally and temporally.

For over 200 miles along the Red Deer River in southern Alberta extensive badland development has occurred. The Red Deer valley, incised from 200 to 350 feet into the prairie surface, formed originally as a meltwater channel draining the Keewatin ice-sheet. Rapid downcutting produced deep valley sides into which active tributary valleys have incised to produce the badlands. These valleys are cut into the almost horizontal, highly erodible claystones and sandstones that comprise the valley walls.

Ten small drainage basins were selected for study on the north bank of the Red Deer near Steveville (Fig. 1). One or two of the basins are dominant, extending well into the upper prairie surface where no headwater competition occurs. As these basins extend they consequently restrict the development of smaller adjacent basins which as a result are currently controlled in form by the steep valley walls of the Red Deer.

The study reveals that an evolutionary sequence of basin form can be traced as the basins enlarge, breach the prairie level, and adopt gully-like characteristics. Form changes are shown in basin hypsometric curves and longitudinal stream profiles. Certain anomolies in the planimetric character of each drainage net can be explained within the de-

Fig. 1 Aerial photograph of the study area



1" to 1000'

signed model of basin evolution.

Much of the current literature devoted to fluvial geomorphology relates basin and channel forms by simple exponential functions. Relatively consistent relationships have been found to exist between increasing basin size and increasing mainstream length, gradient and discharge in a downstream direction. Associated with these changes, channel width and depth increase, width more rapidly than depth. In these studies, hydrologically significant links have been successfully established between two previously independent fields of study, the Hortonian drainage net analysis, and channel hydraulics.

The relationships that hold in a downstream direction in other areas might be expected to hold between adjacent basins in the study area. Hence, the morphometry of the basins may be related to the form of the channel at its watershed emergence for each basin. Detailed measurements of the characteristics of each channel in the field were used to test this hypothesis.

The technique used here is consequently a novel approach to the problem of landform genesis. Although the model itself is applicable only within the confines of the Steveville study area, exponential functions derived compare well with those established for downstream changes in other areas. It is believed that this validates both the model and the techniques used to describe it.

CHAPTER I

INTERBASIN VARIATIONS

The drainage basins in the Steveville badlands have formed, and are forming, within closely similar lithologic environments. Because of this lithologic similarity, and the comparability of climate and processes within the region, a high degree of interbasin morphometric comparability might be assumed to exist. In fact this is not so, and examination of the slope processes reveals that the various erosional mechanisms acting on each of the lithologic units within a vertical sequence produces profound variation in vertical basin form.

The proportional effect of the lithologic units in any one basin differs according to the size and shape of the basin. In explanation of this a hierarchical sequence of basin domination can be outlined, a trend clearly definable in the hypsometric curves and longitudinal stream profiles.

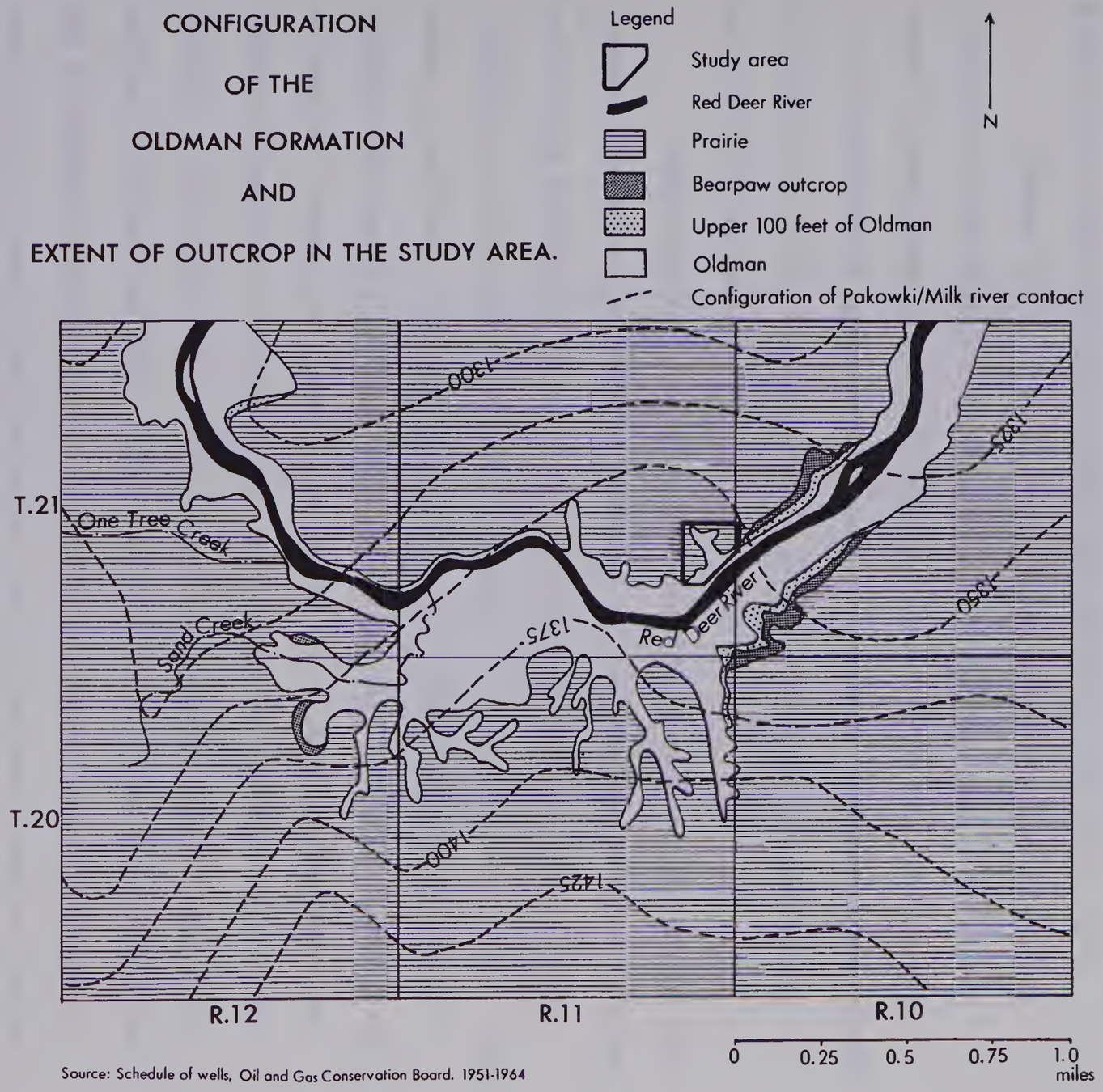
A model is constructed of basin evolution to show that larger basins are dominant and are most active in expanding because they are eroding laterally into the prairie surface free from competition from other basins. Consequently, they impose watershed limitations on smaller less actively expanding drainage systems and thus reduce the overall erosional ability of the smaller units.

A. Lithology

The strata that outcrop in the study area (Fig. 2) comprise 300 feet of the Oldman formation of the southern Alberta late Cretaceous series (Russell and Landes, 1940). These form the second oldest unit of

Fig. 2

CONFIGURATION OF THE OLDMAN FORMATION AND EXTENT OF OUTCROP IN THE STUDY AREA.



Source: Schedule of wells, Oil and Gas Conservation Board. 1951-1964



the Belly River formation within the area. The last 100 feet of the Oldman unit and part of the Bearpaw formation which it underlies are truncated and do not outcrop in the study area. They occur just east of the basins (Fig. 2). The truncated Oldman is overlain directly by 25 to 50 feet of glacial till (Zone A, Fig. 3, see back folder).

Dodson (1969, Fig. 2) constructed a map based on well logs (Oil and Gas Conservation Board, 1951 - 1964) indicating general dip trends in the Oldman formation assuming these follow the configuration of the Pakowki/Milk River contact across the area. The strata are almost horizontal (Fig. 4), dipping only 25 feet per mile to the north-east across the study section.

These strata comprise a conformable sequence of marine and continental shales and sandstones deposited in and around a shallow interior sea in late Cretaceous times. Soft, poorly-bedded argillaceous rocks predominate in the formation and "...deserve the name of clay rather than shale....such beds are commonly covered by a layer of surface wash," (Russell and Landes, 1940, p. 62). Outcrops vary from soft clayey silts to massive, coarse, relatively pure sandstones. Beds of ironstone and indurated sandstone are common, but the thick carbonaceous shales predominant in the Lethbridge area are generally absent in the Steveville outcrop (Williams and Burke, 1964). Fossil vertebrates are terrestrial and freshwater, and noteable in the clean sandstone beds are dinosaur fossils giving name to the Provincial Park on the south side of the Red Deer River.

(i) Mapping

The succession of sandstones and non-fissile claystones that



Fig. 4 Horizontal bedding of strata in the study area.

predominate in the 300 feet high valley walls are important structural units which, since they experience different erosional mechanisms, produce distinctive landform types. In order to ascertain the exact extent of each structural unit between basins, the outcrop geology was mapped in the field. On the basis of Dodson's (1969) map, it was known that the strata would dip only 10 to 15 feet, or almost one contour interval, south-west to north-east across the study area. Thus, once ten sections had been mapped, the outcrop map (Fig. 3) could be constructed by extrapolation along the contours. Locations of the ten stratigraphic sections that were done are marked on Fig. 3 (A - A' ... J - J').

Certain distinctive units aided mapping. In particular, the uppermost thin band of dark carbonaceous shale could be traced by eye across the upper level of all the basins. Beneath this, a band of dark grey non-fissile claystone 14 feet thick overlies 13 feet of clean fine sandstone which, though relatively poorly consolidated, is considerably harder and better cemented than the claystone. Below this, claystone is once again evident and terminates with a very thin band of distinctive sideritic or limonitic concretions of a dark grey to blue metallic appearance which exhibit classical spheroidal weathering where exposed.

Underlying this interbedded claystone and sandstone unit (B) lies a massive outcrop, -90 feet thick- of the clean fine sandstone (C). This is interbedded with thin bands of indurated sandstone. The existence of these has been ascribed to diagenetic cementing by siderite (Dodson, 1969) and hence they vary somewhat in thickness but are laterally continuous and parallel the margins of the outcrop.

Underlying this major unit, more thinly bedded claystone and sandstone strata occur once more (D). Being similar in form to the higher

bands, they are interbedded again with a thin band of ironstone concretions. Beneath these units a very distinctive indurated sandstone band (ss_1) is evident, and in places forms a distinct break in the slope profile (Fig. 5). Three of these bands were found (ss_1, ss_2, and ss_3), varying from 3 to 5 feet in thickness, at elevations of 25 feet, 45 feet and 80 feet above the points of channel emergence from the watershed. Separating these bands is another large unit of clean fine sandstone (E), which extends to 30 feet above the watershed emergence point and is also characterised by bands of ironstone concretions.

Below the third sandstone band (ss_3) another unit of the non-fissile claystone (12 feet) overlies 10 feet of clean fine sandstone which, together with an underlying claystone outcrop form the last major structural zone (F) exposed above the valley floor. In most cases these last two units were overlain by residual deposits and alluvium.

Except for small variations in basin 3, 4 and 5 (whose headwater tributaries have extended only as far as structural zone C), Fig. 3 reveals the lateral homogeneity of the lithology in the study area. Thus each basin is experiencing similar lithologic controls, and therefore no basin should be considered as structurally distinct from any other. The implications of these initial observations will be discussed in section C.

(ii) Comparison with other areas

An examination of these lithologic units reveals a close parallelism with those of other badland areas. The Chadron and Brule formations belonging to the White River group (Oligocene) of South Dakota have been similarly exposed under the action of a semiarid climate to produce



Fig. 5 Structural break caused by indurated sandstone bands in basin 1.

badland topography. The description of the two formations given by Schumm (1956a) near Wall, South Dakota, makes it possible to compare the Chadron formation with the soft, non-fissile claystone units seen at Steveville, and the Brule formation with the clean, fine sandstone units. Smith (1950) has described these two formations as together constituting poorly consolidated clays and silts with some channel sandstones interbedded with nodular clays and calcareous concretions, and emphasises their horizontal bedding.

Schumm (1956a) ascribes the swelling quality of the strata when wet to the presence of the clay minerals illite and montmorillonite. Both of these minerals are present in the Oldman unit (Dodson, 1969). The greater consolidation of the Brule is due to a larger percentage of calcareous cement, up to 10 or 20 per cent as compared with 4 per cent for the Chadron (Smith, 1950). The presence of calcareous cement is readily apparent at Steveville. Some clean sandstone slopes and areas of wash are almost white, and when broken a fine white dust testifies to its calcareous nature.

Similar comparisons can be made with the homogeneous clays of the Chinle formation near Cameron, Arizona, which from their description (Gregory, 1950) appear similar in form and erosional mechanism to the claystones existing in this area. Examples of erosional process studies in these comparable areas will be cited to elucidate landforms seen at Steveville.

B. Landforms and processes

Badland areas are characterised by an extremely fine, highly rilled drainage network separating short steep slopes and narrow inter-

fluves and are often completely free of vegetation (Fairbridge, 1968). Characteristic badland landforms are all apparent in the Steveville area, including hoodoos (Fig. 6), miniature pediments and fans, and buttes and mesas of various forms (McPherson, 1968). These features result from the interaction of the highly erodible bentonitic claystones and sandstones with the semiarid climate, characterised by extremely seasonal rainfall. About 70 per cent of the regional mean annual precipitation of 12 to 16 inches falls between April and September. Thus the landforms can be regarded as a result of the unique evolution of individual slope elements closely controlled by lithology and developing under a closely similar climatic regime.

Much literature has been devoted to studying the various erosional mechanisms acting on particular lithologic units (Schumm, 1956a,b; Campbell, 1970; Smith, 1950). One particular mechanism of erosion that occurs in badland areas is piping. This has been defined by Mears (1963, p. 101) as "...subterranean erosion initiated by percolating waters which remove solid particles from clastic rocks to produce tubular underground conduits." A notable characteristic of all semiarid and arid zones, piping affects all material from clay size to gravel, and in particular the claystones and the other less fissile sediments associated with badlands.

Essentially, piping is a non-solution process occurring in the crowns, slopes and channels where differentially erodible materials are interbedded. Water percolates through the claystone units which have the highest infiltration capacity (Schumm, 1956a) and escapes along gully bottoms or at contact with a less permeable horizon. Thus, tubular conduits are eroded into these sediments, and collapse features and the disappearance of stream channels on the ground surface can result (Fig. 7a and b).



Fig. 6 Effect of thin bands of indurated sandstone in controlling slope form and initiating the development of hoodoos. Note armour of clay ironstone fragments, near left.



Fig. 7a Piping in structural zone D. This is a second order tributary on the eastern flanks of basin 6.



Fig. 7b Surface slumping as a result of diversion of flow through underground conduits.

It might be supposed that piping is deranging the drainage pattern to such an extent as to affect the planimetric form of the streams. There is certainly a possibility that rain falling in one basin emerges by means of these underground conduits into another basin. This, however, is probably a very minimal amount, and major piped areas occur significantly in only one major structural zone D (Figs. 3, 7a). Field investigation revealed no evidence of interbasin piping. Minor drainage derangement can be seen on the eastern side of basin 6, and some streams were difficult to map from the aerial photographs. Field reconnaissance indicated that these streams could be demarcated second order. Consequently, piping was not considered to be a major source of interbasin variations in the planimetric form.

More significant mechanisms controlling slope evolution have been studied by Schumm (1956a). In the Brule and Chadron formations of South Dakota the presence of two distinct topographic forms is obvious. Steep, sharp-crested slopes with high mean maximum slope angles, are developing on the Brule formation and correspond to those developing on the clean fine sandstone units at Steveville, which are heavily rilled and compact when dry. On the interbedded Chadron, similar to the non-fissile claystone units in the study area, rounded rolling interfluves are developing with slope angles generally less than those on the Brule. The surface of the Chadron appears similar to that of the Steveville claystone, being friable and disaggregated when dry (Fig. 8).

Smith (1950) has attributed differences in slope angles in South Dakota formations to the higher content of calcareous materials in the Brule clays. The surface of the two formations is quite different and in description and from photographs appears to be equivalent to the



Fig. 8 Sharp divides on the clean sandstone (L) where heavy rilling testifies to rainwash. Rounded divides of the underlying claystone (R) testifies to creep. East of basin 6 at the fenceline.

two main lithological types found at Steveville.

Schumm (1956a) investigating the same topography, interpreted the differing form of the two lithologic types as a result of two distinct erosional mechanisms. Broad, round interfluves and the loosely disaggregated surface of the claystone might suggest rapid erosion, and yet as Schumm (1956a, p. 700) points out, the Brule formation has retreated far back from the Chadron. He added "...the Chadron thus appears to act as a more resistant rock." As an explanation for this, the higher infiltration capacity of the Chadron was exhibited by pouring a canteen of water onto the friable surface, and this was immediately absorbed, whereas water poured onto the Brule ran off as sheet flow. Schumm, therefore, ascribed the mode of erosion on the claystone as predominantly creep[1], whereas the Brule with its steep slopes and low permeability was eroding more quickly by a process of rainwash[2]. In analysis of these mechanisms, Schumm drew the parallel between the differences in characteristics between humid and arid topographic types, or the ideal forms represented by Davis (1909) and Penck (1953) in their respective cycles of slope evolution.

In an attempt to further document these erosional mechanisms, and to assess the volume of material removed in the summer period in the Steveville badlands, Campbell (1970) established nine runoff plots on a

[1] Creep is defined by Bryan (1922, p. 89) as "the slow movement of soil and rock waste down the slope from which these materials have been derived by weathering", i.e. primarily due to gravity, "but is facilitated by the presence of water, alternate wetting and drying, freezing and thawing, growth and decay of roots, and the work of burrowing animals," (quoted in Schumm, 1956a).

[2] Rainwash is defined by Bryan (1922, p. 89) as, "the water from rain, after it has fallen on the surface of the ground and before it has been concentrated into definite channels," 1922 (quoted in Schumm, 1956a).

random selection of sites in the main study basin (6, see Fig. 3), which represented"... a variety of micro-environmental conditions", (p. 3). In particular, plots 1 and 4 represented erosion rates on a thin band of clean fine sandstone underlying the uppermost indurated sandstone band. These plots had southwestern and western aspects respectively, with slopes of 25° and 12°. The former plot produced the most sediment (3.68 lb./yd.^{2}) representing an average total of 7.59 in. of precipitation in three and a half months, and also yielded the sandier sediment. This was accredited (p. 8) to the greater slope angle on plot 1. Plot 4, at 12° produced only 1.65 lb./yd.^{2}) in the same period. Thus slope was related to erosion rates within the same unit.

The lowest yielding plot, #3 (0.06 lb./yd.^{2}) was on glacial till, and was densely vegetated having a slope of 28°30'; whereas maximum yield occured on plot 8 (10.40 lb./yd.^{2}) which was on the highly fractured unvegetated carbonaceous shale at a similar angle (32°). Thus the degree of vegetation, and differences in lithologic composition resulted in quite different amounts of sediment yield despite similar slope angles.

The second highest producer (7.57 lb./yd.^{2}), plot 5, had the lowest angle (1°) and was positioned on a similar fine sandstone unit as plots 1 and 4, and moreover had a certain amount of protection from clay ironstone fragments. This plot had a far greater yield than plots 6 and 7, (24° and 20° respectively) producing 0.52 and 0.38 lb./yd.^{2} in the same period. This was ascribed (p. 15) to the protective effect of a high percentage of the ironstone fragments on the latter two, especially plot 7. Plot 9, at 5° was situated on the clean sandstone band but produced only 0.83 lb./yd^2 quite different from most plots on the unprotected clean sandstone which in general produce high yields (1, 4 and 5). The one plot on the non-fissile claystone surface (plot 2) at 11°30' had a low

yield. This may be partly due to the vegetation cover but in general is in agreement with the generalisations about rainwash and creep made by Schumm (1956a).

This study (Campbell, 1970) pinpointed the primary control that lithologic types in the area have over both erosion rates and hence slope form, despite the obvious effect that slope angles have in increasing erosion efficiency. Vegetation and ironstone armour were seen as effectively reducing erosion on some slopes.

Hence it can be generally stated that the two primary controlling factors in slope development in the Steveville badlands are lithologic type, and position of the unit within the stratigraphic sequence. The lithologic composition ultimately controls the type of erosion, i.e. whether rainwash or creep will predominate. The position in the stratigraphic sequence is important for two reasons. First, if clay ironstone fragments are adjacent they will tend to armour an otherwise unprotected underlying unit. Second, because the action of a more resistant unit above an outcrop will act as a protective caprock, and below it will act as a temporary grade control, as with the indurated sandstone bands (ss_1, ss_2, and ss_3). The opposite effect is produced by less resistant units above and below. Schumm (1968, in Fairbridge, p. 47) stated this in the following manner;

> "In badland areas, where lithologic differences occur, the regularity of the slope is disturbed, and often miniature cliffs form below the resistant layers, in fact sod can act as a caprock when overlying highly erodible sediments."

Characteristics of the slopes evolving on various structural zones in the Steveville area are generalised in Table I.

In analysing the total sediment yield data from the whole study

TABLE I

Zone	Structural Types Represented	Slopes		Erosion Dominant	Intensity
A	Glacial till with significant vegetation	moderate	General tendency for slope decrease with lower elevation.	creep	low
B	Claystone, non-fissile Clean fine sandstone	moderate high		creep rainwash	low high
C	Clean fine sandstone (a) protected by ironstone (b) enforced by indurated bands	high variable moderate		rainwash rainwash rainwash	high low moderate
D	Claystones, non-fissile Clean fine sandstone (a) protected by ironstone	moderate high variable		creep & piping rainwash & piping rainwash & piping	low high low
E	Clean fine sandstone enforced by indurated bands (s_1, s_2 & s_3)	very high		rainwash	moderate
F	As for B	As for B		As for B	As for B
	Alluvium and residual	low		rainwash & fluvial	variable

period, Campbell (1970) found this to correlate well with rainfall variations in the spring and summer (Fig. 9), and also found the values he obtained to be similar to those of other comparable badland areas. From the point of view of landscape genesis, rates of erosion of this magnitude imply that each slope element must be passing through it evolutionary stages at a rapid rate, but controlled in this by lithology as described above.

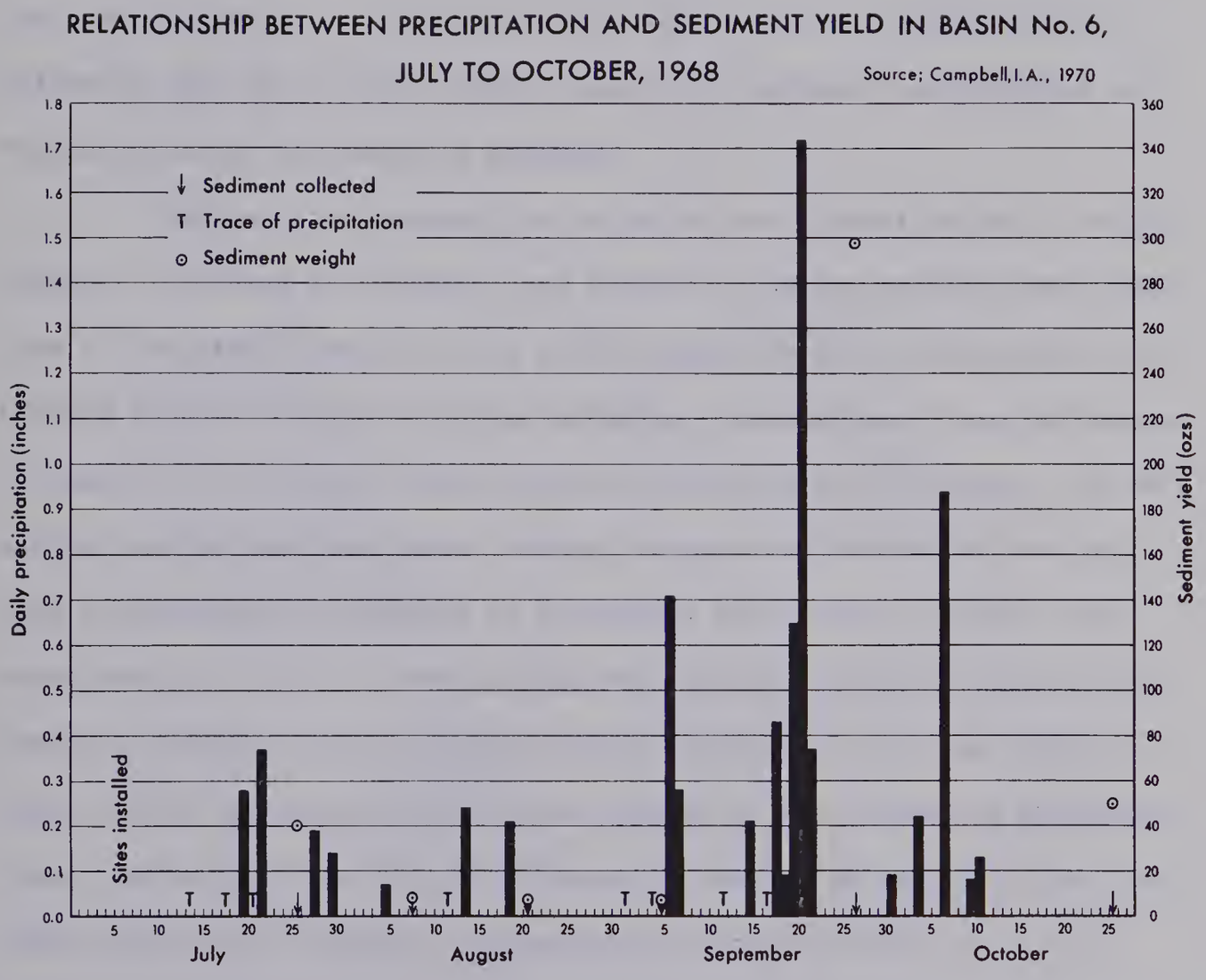
Since lateral lithologic variations are negligible between basins, it might be suggested that total morphology of the landscape, since a combined result of the individual slopes at any point in time, should be identical for each of the ten basins. This is not however, a valid argument. Because, although each slope is evolving according to principles governed by its lithologic and stratigraphic characteristics, in a small basin any one lithologic unit has a much greater proportional effect than in a large basin in controlling basin form. This can be regarded as a function of basin size and shape. A study of the hypsometric curves and longitudinal stream profiles illustrates this effect.

C. Basin Forms

An examination of the study area shows basins of quite different shape and size in close proximity. In addition, some larger basins have extended into the prairie surface, whilst others peripheral to these have made no incision into the prairie level at all, and are apparently developing on a retreating scarp face rising from the alluvial flats to the prairie.

The differences in size can be viewed as a result of interbasin competition during the long-term evolution of the landscape. As a consequence, one or two basins in the study area have adopted more dominant

Fig. 9


RELATIONSHIP BETWEEN PRECIPITATION AND SEDIMENT YIELD IN BASIN No. 6,
JULY TO OCTOBER, 1968
Source; Campbell, I.A., 1970
↓ Sediment collected
T Trace of precipitation
⊙ Sediment weight
Daily precipitation (inches)
Sediment yield (ozs)
Sites installed
July
August
September
October



roles, thus imposing limitations on smaller basins. This process of basin domination can be seen as a "snowball" effect, since a small size advantage in any one basin implies a small but significant increase in the area of headwater activity. Having this advantage, the basin will be more successful when in competition with smaller basins that have poorer headwater development.[3] Where no headwater competition occurs, i.e. when basins have increased in size sufficiently to allow lateral extension into open prairie above, growth is completely uninhibited and the dominance of the basin is assured.

Earlier, the observation was made that classification of slope elements according to lithology and process in no way explains the gross form of individual basins, since smaller basins are more obviously controlled in their form by vertical lithologic variations. This difference in degree of lithologic control can be accredited to two factors. First, smaller basins have less water passing through the system and are less able to eliminate the effects of lithologic differences. Second, and more important, Fig. 10 reveals that the smaller, inhibited basins that have not extended into the upper prairie (example a) are controlled in gross form by geological differences exposed on the retreating escarpment, drawn theoretically as A' - A". Basins now extending laterally into the upper prairie (b), however, are dependent on both the form of A - A' and A' - A". Once the basin has established itself in the prairie, its mode of development changes, it is free to expand and vertical lithology becomes less significant. This process is shown in an examination of the basin hypsometry and the longitudinal stream profiles.

[3]This has been fully explained by Leopold, et al., 1964, p. 415

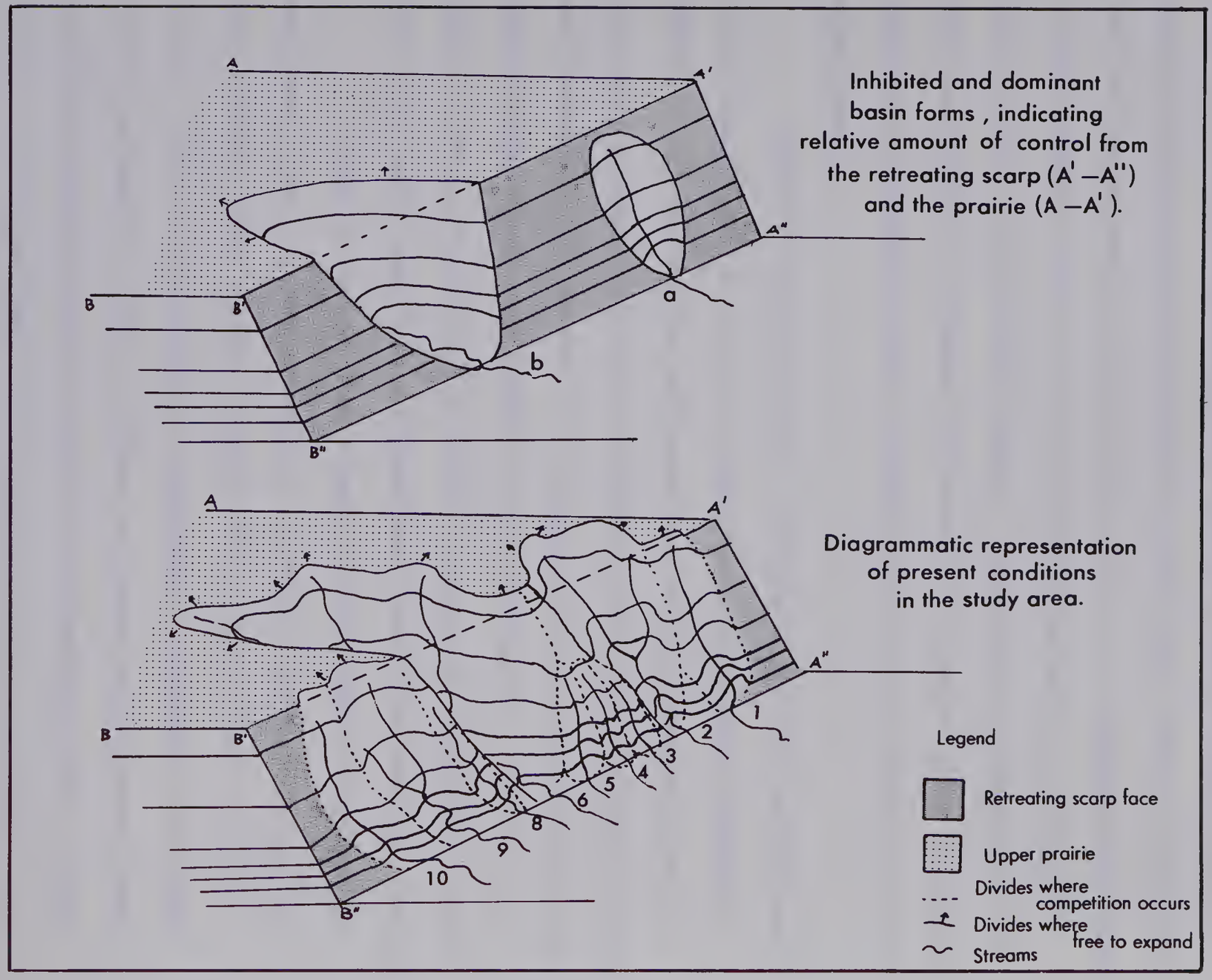

A
A'
A"
B
B'
B"
a
b
Inhibited and dominant basin forms , indicating relative amount of control from the retreating scarp (A' —A") and the prairie (A —A').
Diagrammatic representation of present conditions in the study area.
1
2
3
4
5
6
8
9
10
Legend
Retreating scarp face
Upper prairie
Divides where competition occurs
Divides where free to expand
Streams

Fig. 10



(ii) Hypsometry

Different types of hypsometric curves have been devised for a variety of uses. Most widely used is that of Strahler (1952b) which relates the percentage area enclosed by each contour above a given datum (a/A) to its percentage elevation (h/H). Due to the dimensionless nature of this technique comparison is possible between basins of different sizes within a two-dimensional figure. Differences in sinuosity of the curve led Strahler to propose a three-variable function which adequately described all possible curves, and these functions were given geomorphic significance.

Within this square figure, "youthful", "mature" and "old age" curves were described. The use of this terminology stems from Davis' (1909) concept of landform interpretation, in which considerations of structure, process and stage of development were said to totally describe any landscape. This framework is now regarded as too restricted in vision, incorporating rather unconformable concepts of "grade" as outlined by Gilbert (1880) and later Mackin (1948), and also the idea of any equilibrium which might involve time-independent states other than the peneplain.

Within an open system framework the concept of "entropy" has been found to be a suitable remedy for these shortcomings. Using this idea, Strahler (1950, 1952a), Culling (1957), Hack (1960), Hack and Goodlett (1960), Chorley (1962a), Scheidegger and Langbein (1962), and Leopold and Langbein (1962) introduced to geomorphology a more liberal conceptual framework than was fostered by the Davisian school. Consequently, certain geomorphic features can be seen in a new light as partially or largely time-independent adjustments. The terms "youth", "maturity" and "old age" are still used, but in a different sense; to attest to level of dynamic equilibrium attained. The percentage hypso-

metric curve has been considered evidence of such levels of equilibrium (Strahler, 1952b).

Since all the basins in the study area are affected by the same lithology and erosional mechanisms, and also can be assumed to be the same age, they should by this argument have identical hypsometric curves. The mode of basin evolution presented above, however, exposed this line of argument as invalid. Differential growth of basins within the constraints of two controlling facets (Fig. 10) has produced basins of different size and form, and in the light of this model the concept of evolutionary stage of development is greatly complicated.

In the system of basins in the study area, the internal energy available within each basin is variously used. Either it is used to compete with other basins, or to expand laterally without competition, or to modify internal form. The nature of the modification process (Section B) is the same in each basin, but the percentage of energy used to this purpose is not proportionately the same for each basin due to varying degrees of lateral expansion activity and interbasin competition.

As a result, basins evolving in A' - A" develop quite different forms from those that have breached the prairie surface and so examining the hypsometric curves for evidence of an "equilibrium" level for the whole area is likely to be quite misleading.

Two other reservations about Strahler's hypsometric integral arise from his statement that the area below the curve represents the volume of land remaining for removal as a percentage of that originally present. Implicit in this is the assumption that all basins have been eroded into an initially flat surface. In the Steveville area erosion has progressed from two controlling facets, the smaller basins developing

relative to a retreating scarp face until incision into the prairie is attained. At this point development is in reference to two planes, the scarp face, and more significantly, into the upper prairie. Thus the square in which the hypsometric curves are drawn cannot in any basin be assumed to represent the total volume available for removal.

A second reservation is that Strahler assumes that watersheds are relatively static features of the landscape and that basins evolve within the watershed in the third dimension rather than extending rapidly in two dimensions. In the study area the choice of watershed posed particular problems. The smaller inhibited basins have well-defined divides, but basins 6 and 2 are extending in a gully fashion into the prairie. Strahler (1952b, p. 1129) uses a basin very similar to basin 6 in illustrating "youthful" bulge, but the "bulge" occurs only because he takes three spot heights on an upper terrace surface and arbitrarily demarcates a watershed between these points. It was not possible to choose any other line than the outline of the basin proper as the watershed for basins 6 and 2 in the study area. There were no drainage lines or contours on the upper surface on the map, and in the field the upper prairie adjacent to basins 6 and 2 could be seen to be making minimal contribution to these basins.

These several problems prompted a different approach to form description than the hypsometric integral in order to discuss basin evolution within the model illustrated in Fig. 10. For each basin the area enclosed by each contour was plotted against the height of the contour. This makes no assumptions about volume of removal and the choice of watershed is not an important part of the construction of the figure; yet it gives a good two-dimensional impression of the distribution of

area with height in each basin. Basin curves were constructed on the same diagram, and synchronised at their emergence point from the watershed, thus occupying their true lateral and vertical positions (Fig. 11).

From the diagram the dominant basins 1, 9, 2 and 6 form a sequence which can be regarded as the mode of basin evolution once the prairie level has been breached. This sequence is quite different from that represented by the basins developing on the scarp face (8,3,4,5 and 10) which again form a size sequence and can be regarded as the mode of basin evolution if the prairie level has not been attained. This latter sequence tends towards upconcavity with a corresponding decrease in structural control. The former tends towards upconvexity, thus clarifying further the theoretical model described by Fig. 10.

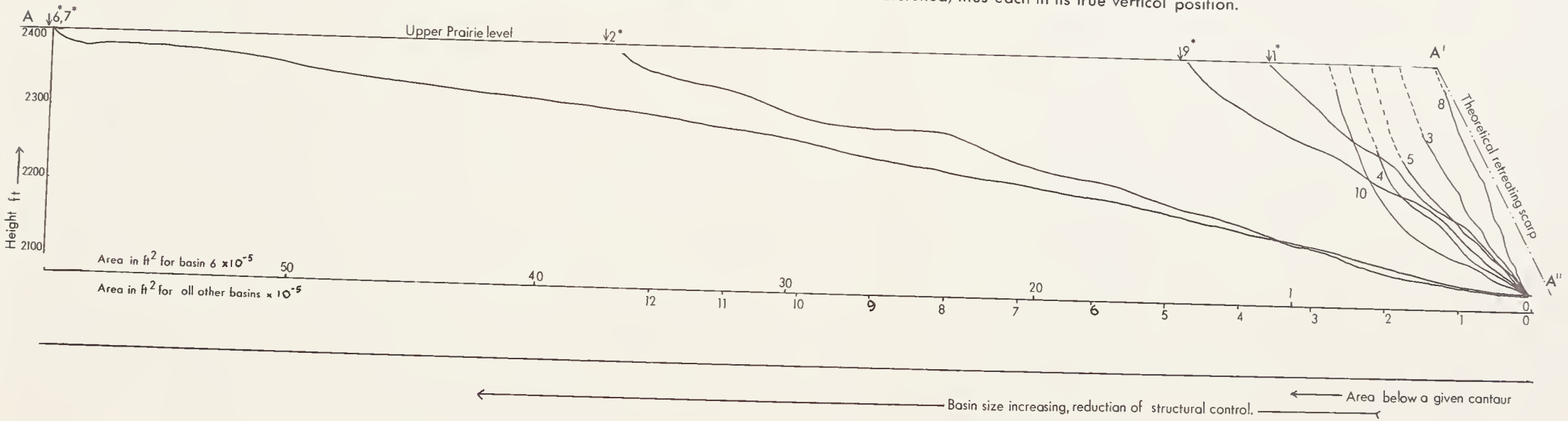
(ii) Longitudinal Stream Profiles

The classical concept of "grade" as defined by Mackin[4] has led some workers to suggest that a graded stream is one which is neither eroding nor depositing along its course, i.e. a stream which will present a completely smooth, concave-up form along its longitudinal profile. However, other researchers have revealed that this concept of the graded stream may be erroneous, since as discharge fluctuates daily, so does the cross-sectional area of the channel and the channel course in response to this. Wolman (1955, p. 47) suggests that the downstream curves on Brandy-

[4] Mackin (1948, p. 464) defined a graded stream as ..."one in which, over a period of years, slope is delicately adjusted to provide, with available discharge and prevailing channel characteristics, just the velocity required for the transportation of the load supplied from the drainage basin. The graded stream is a system in equilibrium, its diagnostic characteristic being that any change in the controlling factors will cause a displacement of the equilibrium in a direction that will tend to absorb the effect of the change."

Fig. 11

MODE OF EVOLUTION OF BASIN FORM. Synchronised hypsometric curves for eoch basin.
Basins synchronised at their point of emergence from the watershed, thus each in its true verticol position.

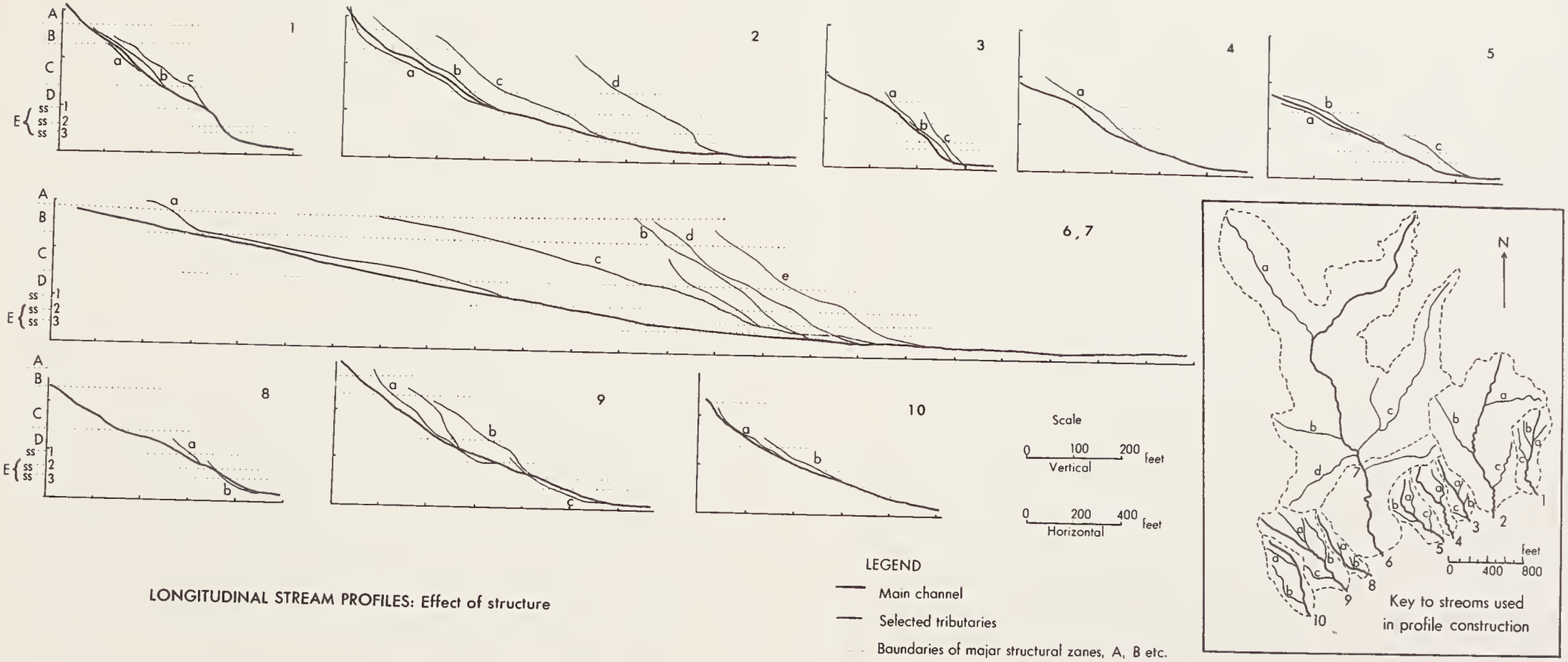


wine Creek imply an adjustment of the cross-sectional area of the channel equally as important as the adjustment of the long profile, and "...there is no way one could predict that the effect of a change in the independent controls would be absorbed by a change in slope rather than a change in cross-section." Thus, it may be that a stream can be virtually always adjusted (Hack, 1960, p. 85) in the sense of being "...graded or in a steady state without necessarily presenting the smooth longitudinal profile." (Chorley, 1962, B. 4).

As a result of these problems inherent in the grade concept, and following the discussion of the hypsometry, it was concluded that a search for a graded or equilibrium state in the long profiles would be equally misleading as the use of the hypsometric integral. Rather, a sequence of stages is sought which will reveal the evolution of stream profiles in inhibited and dominant basins respectively. Analysis of Fig. 12 shows a series of long profile forms that corresponds well with the forms represented by the hypsometric analysis.

Profiles 8,3,4,5 and 10 represent a sequence of profiles as size decreases for basins on the retreating scarp. Structural control is apparent in the forms, particularly the indurated sandstone bands that outcrop at 30 feet, 50 feet, and 80 feet above the valley floors. Basins 1,9,2, and 6 similarly represent a size sequence for basins which have to various degrees breached the prairie surface, and a gradual decrease in gradient and elimination of structural control is quite apparent. All streams were plotted together as with the hypsometric curves (Fig. 13) to illustrate the sequence, and Fig. 14 is a plot of gradient change with increasing basin size. This last graph illustrates very clearly the uniform trend of gradient decrease once the prairie level has been breached, whereas the inhibited basins have no discernable trend, being variously

Fig. 12



LONGITUDINAL STREAM PROFILES: Effect of structure

Fig. 13

MODE OF STREAM PROFILE EVOLUTION. Diagrammatic representation of the lowering of profile gradients with increasing basin size.

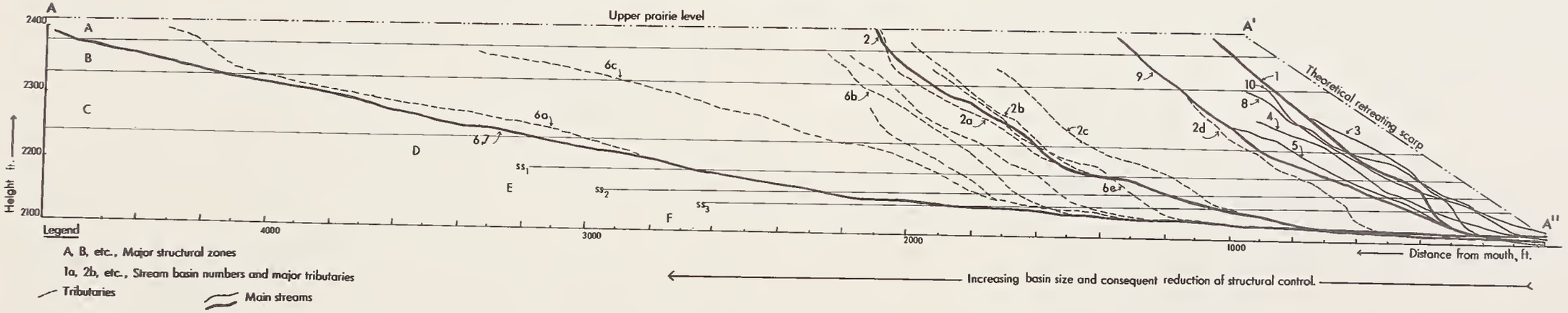
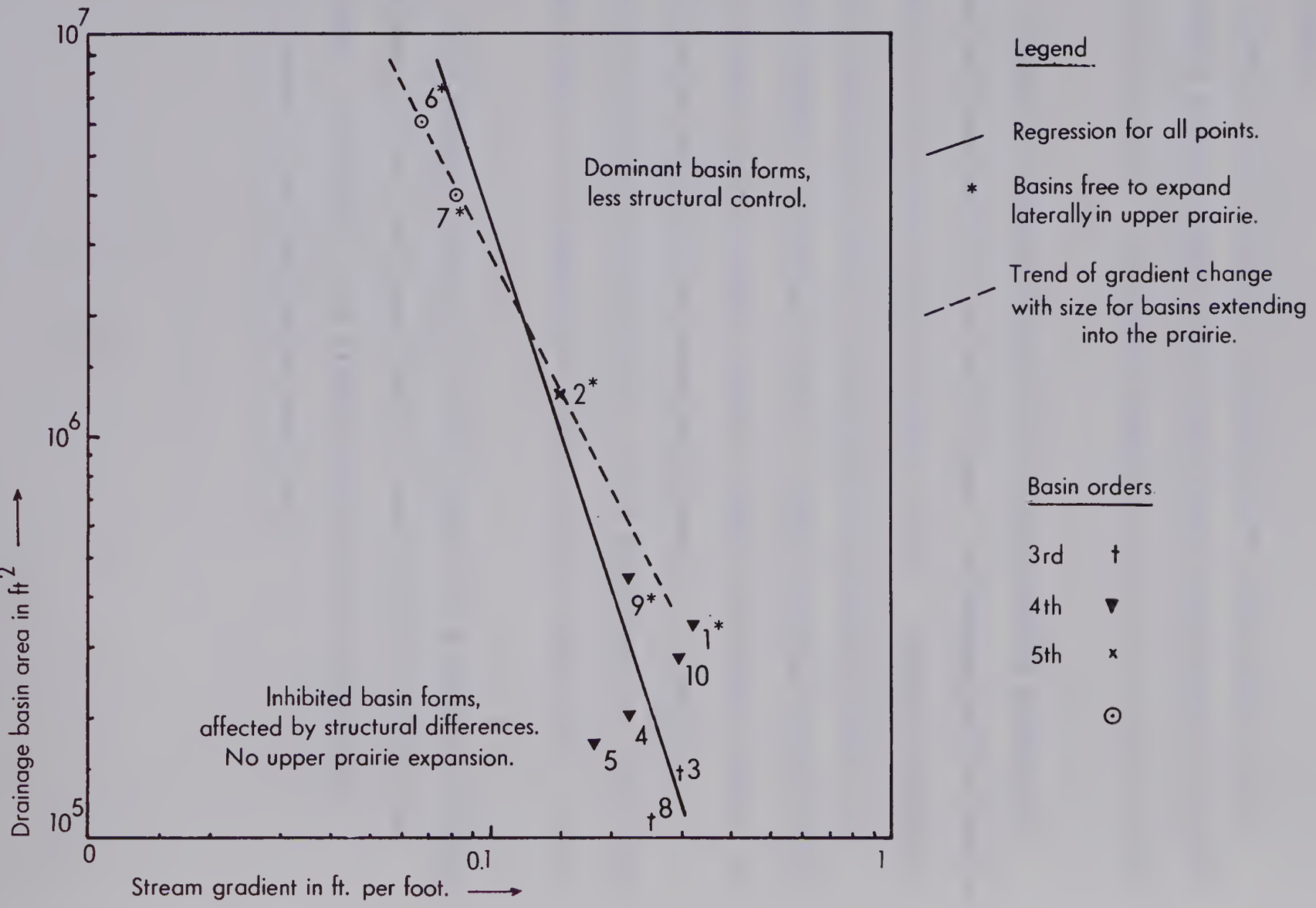


Fig. 14 **EFFECT OF BASIN SIZE AND PRAIRIE EXTENSION ON STREAM GRADIENT AND REDUCTION OF STRUCTURAL CONTROL**





steep and controlled by structure.

D. Conclusions

Lithology and erosional mechanisms have been shown to vary insignificantly between basins, each slope being controlled in its evolution by the structural type and the position of the unit in the vertical sequence.

A model of basin development has been presented wherein all basin forms are controlled by two main facets, the theoretical retreating scarp face (A' - A") and the upper prairie (A - A'). Basins which have not breached the prairie level are primarily controlled in form by A' - A" and the lithologic variations therein. Basins that have breached the prairie level are free to expand, dominate smaller restricted basins, and are controlled in their mode of evolution by both A - A' and A' - A", these tending towards a gully-like form. This model has been illustrated by an examination of the hypsometric curves and the longitudinal stream profiles.

The following chapter (II) will examine the planimetric aspects of the drainage system, and consistency in areal division within the network sought. Deviations from fitted regression lines may be explicable in terms of the conclusions made here.

CHAPTER II

THE MORPHOMETRIC ANALYSIS OF THE BASINS

Morphometric techniques provide geomorphology with an extremely useful analytical tool, especially where rigorous fieldwork is used as a supporting adjunct. Despite theoretical objections to some drainage network classificatory techniques, morphometric ratios and deviations from fitted regression lines in particular are useful indicators of drainage net development. Hortonian stream network analysis and its application in the study area will be described in this chapter, and consequently further light can be shed on the evolutionary model presented in Chapter I.

A. Theoretical considerations.

Morphometric analysis, or morphometry, is the measurement and mathematical analysis of the configuration of the earth's surface (Penning-Rowsell, 1969, p. 1). Ideally, the technique should be regarded as merely the first three stages in a four-fold research process; classification of forms, observations and measurements; analysis of data within the model; and prediction with other geomorphic variable systems (synthesis). The ability of the model to connect with other process systems is not always fully realised, and morphometry has often been regarded as a merely classificatory or at the most, comparative technique.

(i) Classification

The choice of a suitable classificatory technique for drainage nets has proved problematical. Horton (1945), from whose work most basin morphometry stems, devised a geometric-series model for drainage

networks and derived two laws of drainage composition. The method of classification of the segments was rather complicated. Unbranched tributaries were assigned the order 1, unless they join, in which case the predominant tributary (longest - or nearest in a line with its continuation) and its continuing part became order 2. This was repeated until all stream segments were consecutively numbered.

Some workers (Leopold and Miller, 1956; Brush, 1961; Leopold and Langbein, 1962) followed Horton's classification method. Others, (Schumm, 1956; Melton, 1957; Coates, 1958; Maxwell, 1960: Smith, 1958; and Morisawa, 1962) used Strahler's (1952b) modified system to avoid the subjective choice of a predominant tributary (see Shreve, 1966, p. 18). In Strahler's system, the first order segments are enumerated 1. A second order segment is formed by the joining of any two first order streams, and so on. During analysis Horton and Strahler methods produce slightly different results, so that it becomes mandatory to define the system adopted for study.

Although there exists a basic need for a scheme of network subdivision, the usefulness of Horton's (1945) hierarchical system and Strahler's (1952a) modified system has been questioned on both classificatory and analytical grounds. Relevant here is the criticism that a fourth order basin is not necessarily twice as large, nor does it necessarily have twice as many first order streams, as a second order basin, i.e., the scale is an ordinal one (Penning-Rowsell, 1969), and the streams are artificially ranked and do not represent equal intervals on some continuous scale (Krumbein, 1958). Also, these rank intervals appear too large, so that two third order basins, for example, may possess very different numbers of second order streams and hence may be of very

different sizes. This was revealed by Shreve (1966) in his study of 246 natural networks for which the order 3 had been assigned using Horton and Strahler. Histograms of variations in bifurcation ratio revealed wide divergences. Shreve (1967) proposed a system of ordering based on probability theory in which all streams were assigned a "magnitude" according to the number of first order streams they ultimately drain. Networks of equal magnitude possess the same number of links and forks and are "...therefore comparable in topological complexity,...a reasonable conjecture is that they are comparable in other ways as well." (p. 178). This, however, has yet to be proved.

Shreve's (1967) ordering technique is equivalent to that of Scheidegger (1965), which is also related to the idea of magnitude. According to Scheidegger, this technique avoided the problem that one small first order link could cause an increase in the order of the main channel, an error inherent in Horton's classificatory system. However, Shreve (1967) points out that the lowering of the basin magnitude by 1 in the Scheidegger system causes first order streams to become "non-existent".

Strahler's (1952a) method is adopted here for the following reasons. First, the ordering system is easier and less subjective than that of Horton, (1945) and as Scheidegger (1968, p. 1015) suggests and also explains mathematically, "Horton's laws can also be expressed in the same fashion for Strahler stream segments as for Horton streams without making any significant error." Second, the geometric-series analogy allows regression analysis, which the Shreve (1967) and Scheidegger (1965) systems do not do and which is advantageous even if only for clarity of comparison. Third, despite the theoretical problems involved in analysis (see below), the technique has been frequently used and is well-documented.

(ii) Observation and Measurement

Measurements can be made from the network when all the orders have been assigned according to one system or another. These fall into three types; linear, planimetric, and those involving elevation differences. A complete table of these measurable properties has been produced by Strahler (1958).

Much of this data can be derived from maps. For the ten basins used in this study, length, number, and area drained by each stream of each order was ascertained from an enlarged contour map of the area (1" to 200') which had drainage lines that were visible on a stereopair of aerial photographs (1" to 1000') transferred onto it. The total population was used in all cases except for the average drainage area of first order basins in basin #6, for which a ten per cent random grid sample was taken. These were the only data considered significant for analysis (see Chapter I and below).

Other data may be derived from a combination of factors, such as drainage density ($Dd = \Sigma L/A$), Horton (1945), and relief ratio ($r_H = H/L_B$) Schumm (1956b). Consequently not all morphometric measurements are necessarily dependent on order or magnitude.

As Penning-Rowsell (1969) has said, although many of these indices are over ten years old, none can be satisfactorily be assessed due to the slow rate with which they have been adopted by other geomorphologists, or due to the fact that they have been developed for specific, often empirical uses. Strahler's (1958) ruggedness number has been adopted only in two major subsequent studies - Melton (1958a and b) and Lustig (1965), and it may prove interesting in the future to compare this with Smith's (1950) texture ratio, or Schumm's (1956) relief ratio.

(iii) Analysis

Correlation of properties is conducted either by comparison of means, by regression and correlation techniques, or by multivariate analysis. Apart from the simple comparison of means, which has limited significance (for example, Horton's comparison of Dd for two drainage basins) most morphometric studies have relied on regression analysis. The use of the regression technique stems immediately from Horton's law of stream numbers[1] and his law of stream lengths[2]. The former can be expressed as $N_o = r_b^{(s-o)}$ from which it follows that Log. $N_o \alpha O$ and $r_b = N_o/N_{o+1}$, or the antilogarithm of the gradient of the log/normal regression Log $N_o \alpha O$. The latter law may be expressed as $\bar{L}_o = \bar{L}_1 r_L^{o-1}$ from which it follows that Log. $\bar{L}_o \alpha O$ and $r_L = \bar{L}_o/\bar{L}_{o-1}$, or the antilogarithm of the gradient of the log/normal regression $L_o \alpha O$. Similarly, Horton implied a third law from the suggestion that the total length of streams of a given order is the product of the number of streams and the length per stream. N_o is dependent on r_b and increases with O, whilst L_o is dependent on r_L and decreases with increasing stream order. Thus the expression $L_o = L_1 r_b^{s-o} r_L^{o-1}$ should, when L_o is regressed with O, have either a maximum or a minimum for some particular stream order, although this may not occur in all cases.

1 This states that, "the numbers of streams of different orders in a given drainage basin tend closely to approximate an inverse geometric series in which the first term is unity and the ratio is the bifurcation ratio" (Horton, 1945, p. 291).

2 This states that, "the average length of streams of each of the different orders in a drainage basin tend closely to approximate a direct geometric series in which the first term is the average length of streams of the first order," (Horton, 1945, p. 291). The ratio here is the stream length ratio, r_L.

Horton (1945, p. 292) stresses the importance of the ratio ($r_{bL} = r_b/r_L$) and states that it is controlled by "...hydrologic, physiographic, cultural and geologic factors," which "...determine the ultimate degree of drainage development in a given drainage basin." Strahler's modified system (1952a) can be similarly explained in mathematical terms, the bifurcation and stream-length ratios having similar significance but slightly different values (Shreve, 1967) which are still comparable with Horton's r_b and r_L.

Schumm (1956b) proposed a fourth law in which average areas drained by any stream order are approximated to a direct geometric series, i.e. $\log \overline{A}_o \propto O$, and gives rise to a basin area ratio $r_A = \overline{A}_o/\overline{A}_{o-1}$. Many similar ratios have been devised since (see Strahler's [1958] table, p. 280).

This type of analysis raises several problems. It is desirable when sampling to use the whole population (see above for the present study) since if a sample has been used it is usually necessary to carry out a series of tests as outline by Strahler (1954) to determine the frequency distribution of the data and to assess if the sample is suitably representative of the total population. This is particularly problematical in geomorphology as the researcher is ignorant of the frequency distribution of the total population[3].

The analysis, however, raises further significant problems. Shreve's (1966, p. 17) objections to Horton are that the consistency exhibited by the Hortonian method is merely a reflection of the statistics

[3] An exception here is ruggedness numbers, see Strahler's frequency analysis on the Ashdown and Weald Clay outcrops, England. (Strahler, 1958.)

of a "...large number of randomly merging stream channels", (Horton's term) according to the laws of chance "...in somewhat the same fashion that the law of perfect gases arises from the statistics of a large number of colliding gas molecules." Moreover, implied in the definition of stream order is the corollary that no deranged network of any kind can differ far from Horton's geometric law; bifurcation and stream length ratios, and deviations from the straight line regression are significant only in that they imply a sample size too small for generalisation.

Therefore the question necessarily arises what possible purpose the analysis thereby possesses. Smith (1958, p. 1001) comments that the Horton-type regressions he derived from morphometric analysis in badlands of South Dakota "...suggest an orderly development of dendritic drainage in the relatively homogeneous badland clay," but Penning-Rowsell (1969, p. 8) concludes that this statement "...ceases to have any real meaning in the light of Shreve's comments." At the same time he questions the validity of regressing a ranked, ordinal scale with a continuous interval scale.

Here it is necessary to emphasise that even if all branching systems in nature tend to regress in the prescribed Hortonian manner, Shreve's (1967) "probability function" theory cannot be altogether valid. If it were, then every stream on the earth's surface would be just part of one huge, essentially infinite regression line representing the combination of what Shreve (1967, p. 178) describes as "...individual networks [which are] ... part of far larger, essentially infinite networks."

Streams studied on the same scale do not have the same bifurcation, stream length, or basin-area ratios. These differences must be

explained in terms of geomorphic differences.

Horton (1945, p. 290) states that "...the bifurcation ratio is generally higher for hilly, well-dissected drainage basins than for rolling basins," implying that bifurcation ratio (and, presumably the other ratios) are indicative of structure and geomorphic stage, and even Milton (1966, p. 95) rather reluctantly commented that the bifurcation ratio "...is at least partly controlled by geomorphic factors."

Examples of the usefulness of morphometric analysis are well documented in the literature. For example, Chorley and Morgan (1962) in a comparative study between two basins, one in the U.S.A. and one in England of similar size and shape and studied on the same scale revealed differences in drainage density and bifurcation. They attempted to ascertain,

> "...to what extent one can explain the differences in regional morphometry by reference to existing conditions, and if it is possible reasonably to extrapolate the gross differences in present erosional mechanisms back in time." (p. 28.)

Both these objectives were successfully met, the morphological differences being accredited to differing rainfall intensities since the Miocene!

Schumm's (1956b) small scale model study of morphometry and other characteristics of a fifth order basin in the badlands at Perth Amboy, New Jersey, is an outstanding example of the application of the Strahler method as an explanatory tool and comparative technique. In discussing the use of the Hortonian laws, Schumm (p. 602) has said "...stream order analysis permits comparison of the drainage net developed on the Perth Amboy terrace with patterns developing under natural conditions," and uses the morphometry of two other basins Chileno Canyon, California, and Mill Dam Run, Maryland, as comparisons. No fundamental dissimilarity was

found to exist between the model basin and the other two for Horton's first law, but in analysis of total stream lengths, up-concavity of the regression lines for Chileno Canyon and Mill Dam Run basin was ascribed to youthful channel development. Schumm (p. 604) states that,

> "...continued channel development might be expected to raise the value of the fifth order length, bringing it closer to the fitted regression line ... and ... shortness of the fifth order segment at Perth Amboy may be due to the truncation of the drainage pattern..."

Very useful comparisons such as this, where deviations from a fitted regression are given geomorphic significance, would be completely discredited by Shreve (1966), and yet the value of Schumm's work is unquestionable.

Leopold and Miller (1956), attempting to establish relationships between the drainage net and channel hydraulics for ephemeral channels in New Mexico, used Strahler analysis. They discovered a high bifurcation ratio which they ascribed to the nature of a heavy rilled network development.

Explanations of differences in bifurcation, length, and area ratios, and deviations from the prescribed Hortonian lines in terms of the model presented in Chapter I (C) were attempted for the ten basins examined for this study.

(iv) Multivariate Analysis and Connections with other Variable Systems (Synthesis)

Comprehensive geomorphic studies in the future will no doubt become increasingly dependent on sophisticated multifactor and systems analyses for data interpretation. The continued isolation of two variables at a time for study (i.e. regression techniques) can only approximate one

link in an interrelated causal system.

Melton (1957) used the synthetic qualities of the morphometric set by successfully relating variables of morphometry derived from 1: 24,000 U.S.G.S. maps to a surficial variable set and a climatic variable set. Using cluster analysis on a large amount of morphometric and field data, he found that the isolated connections helped to explain more closely in totality "...the true nature of causal relationships in a basin". (Melton, 1958a, p. 493). Melton's (1957, 1958a & b) models are outstanding in their predictive ability.

Morisawa (1959), too, used multiple regression and correlation to relate a morphometric variable set with a hydrological variable set simultaneously. Factor analysis (where many associated variables are combined to produce fewer factors) has been used by the Tennessee Valley Authority (1965, p. 26) to show how "...morphometric data have great potential for substitutes of hydrological information." The serious problems in this technique have been described by Carson (1966) who suggests as a remedy the use of cluster analysis such as Melton (1957) used.

A similar morphometric synthesis has been demonstrated by Strahler (1958). Using maps he attempted to solve process and form inter-relationships using dimensional analysis in which all variables are reduced to combinations of mass, time, and length components. Variations in surface characteristics were explained through drainage density changes within a process/response model. This model again supports the suggestion that geometric similarity of basins can be explained in terms of similar physical environments.

However, complete data on all aspects of basin parameters are

required for these types of multifactor analyses. In the absence of total documentation of an area, the selection of significant causal factors, found to be related in other studies, can adequately pinpoint key connections between the morphometric variable set and, say, the channel geometry set.

Brush (1961, p. 5) attempted to relate hydraulic, morphometric and geologic conditions of selected streams in central Pennsylvania. In discussing the use of the Hortonian system in drainage net analysis he suggested restrictions were evident because "...it is often difficult to evaluate the significance of the parameters: mean length of streams and mean size of drainage area." Brush proposed to use, as Hack (1957) had done, the actual length of the main channel L_s from the drainage divide as a parameter of drainage area, A_s and to relate this last variable to channel geometry. However, it is apparent that total length of main order stream is the last point on the Hortonian regression of total (or average) stream length[4] with order, and that the total basin area is the last point on the regression of average area drained with order. Thus this intermediate relationship (L_s with A_s) is already implicit in Hortonian analysis and the use of the complete Hortonian analysis will enhance any such study since it illustrates how the relationship L_s with A_s is carried through the drainage net into each basin of each order. Also, if A_s and L_s are found to be related to the channel geometry variables, then it should be possible to predict how the geometry parameters vary within the drainage net by extrapolation. The ten basins in the Steveville badlands were studied with the object of

[4] In using Strahler this is not exactly the case. The theory still holds, however, but the main order channel length must be extrapolated up to the divide.

establishing the same connections both within and between basins.

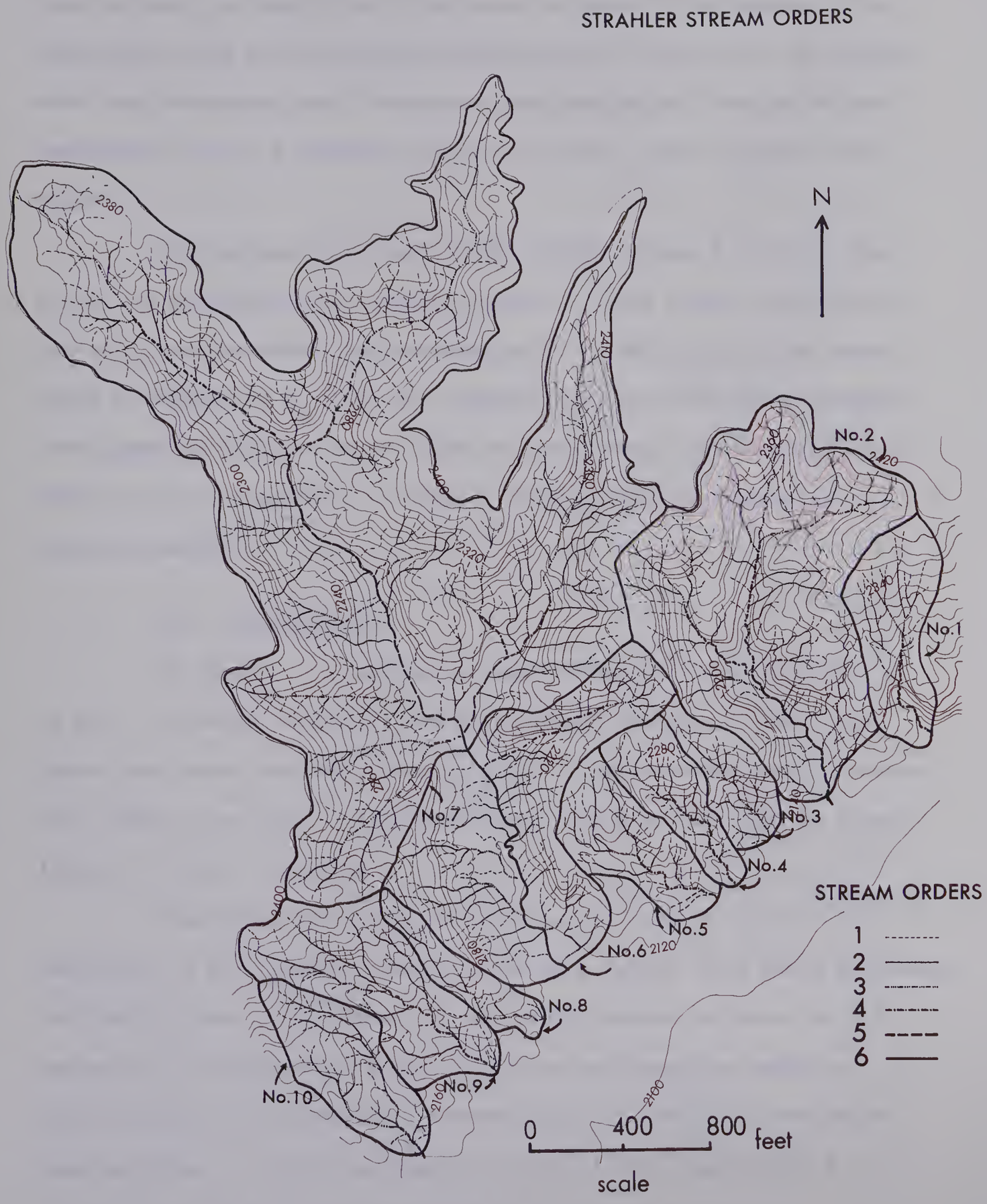
B. Morphometric Analysis of the Steveville Area

Morphometric analysis of the ten drainage basins was based on the photogrammetric contour map (1" to 400') produced by the Technical Division, Department of Lands and Forests, Alberta, for which the accuracy of the 20' contours has been given as plus or minus half the contour interval. The map base photographs were enlarged to be at the same scale on the map, and were viewed stereoscopically with a transparent overlay. All visible drainage channels were traced and transferred to the contour map.

A good knowledge of the field area from the ground was found to be invaluable in identifying the channels. Care was taken to be as consistent as possible in enumerating first order streams. The area is excessively rilled, and what is probably a third or fourth order stream in actuality was demarcated as first order, this being the smallest permanent channel that could be identified on the photographs.

The map was enlarged to twice its original size, and the streams ordered according to Strahler's (1952) method (Fig. 15). At this scale, (1" to 200'), number, total length, average length, and average area drained for each order stream within each of the basins was determined. (Appendix A). Then the following regression lines were constructed; Log. N_o with O, log. L_o with O, log $\bar{L}_o$ with O, and log $\bar{A}_o$ with O, for each basin in turn. The data for N_o, L_o, $\bar{L}_o$ and $\bar{A}_o$ for all the basins were regressed with O on four summary graphs (Fig. 16-19) for which a best fit line for all points had been constructed by the method of least squares, so that basins could be easily compared.

Fig. 15


STRAHLER STREAM ORDERS
N
No.1
No.2
No.3
No.4
No.5
No.6
No.7
No.8
No.9
No.10
2380
2300
2400
2320
2340
2360
2280
2420
2440
2200
2180
2120
2100
STREAM ORDERS
1
2
3
4
5
6
0
400
800
feet
scale



For each basin, bifurcation, stream length, and area ratios were derived, and variations illustrated by means of histograms and by correlation with basin elongation ratios (Figs. 20 and 21). The regressions and histograms were interpreted, and deviations from the fitted regression line, and between morphometric ratios, were variously explained.

The last point on each of the graphs on Log. $\overline{A}_o$ with O, (i.e. A_s, s, for each basin) was used to regress A_s with s (Fig. 22) and similarly L_s was determined and regressed with s (Fig. 23). L_s was considered a function of A_s (Fig. 24). Generalisations about the geomorphic development of the area in the light of conclusions from Chapter I were made from these analyses. In Chapter IV these will be related to channel geometry.

(i) Interpretation

In Chapter I a sequence of basin domination was described. In this, the basin hypsometry and longitudinal profiles of the larger basins that have breached the prairie level form a developmental sequence that differs from that of inhibited basins which are not free to expand laterally without competition.

Planimetrically these differences are evident, basin 6 in particular is more angular, having taken on a "gully" form since breaching the prairie level. Consistency in the areal network division for all basins can be exhibited by using the Hortonian regression analyses. Basins inhibited in their form, however, exhibit consistent deviations from the lines. A close examination of the plotted regressions N_o, L_o, $\overline{L}_o$ and $\overline{A}_o$ with O for each basin help to illustrate this point.

Fig. 16 shows the log/normal regression of N_o with O for all the basins in the area. All the data were used to construct the thick black line, the best fit line to all points. Points used to construct this line that belonged to the same basin were joined so that all basins are represented in one figure, and each basin can be compared to the mean regression line. Theoretically, regression lines for all basins of the same order, s, should coincide along their length, and all basin lines should intersect the abscissa at s. Because all basins of the same order are not the same size in this case, there are divergences in the number of first order streams within basins of that order, and correspondingly the lines make different intercepts on the ordinate. The largest basin should have the highest value ordinate intercept, etc., and so on for each basin according to the geometric progression theory; having slightly different bifurcation ratios but still intercepting the abscissa at s.

A similar explanation for the construction of Fig. 17 can be given. The thick black line is the best fit to all points, although in this case the basin lines do not intercept the abscissa at any particular point. Theoretically the uppermost line should represent the largest basin (since the total length of streams of any given order will be greater the larger the basin) sixth order basins being uppermost and third order basins at the bottom.

Figs. 18 and 19 were similarly constructed, but since the ordinate represents the logarithm of mean values ($\overline{L}_o$ and $\overline{A}_o$ respectively) the position of the individual basin lines is not dependent on the ultimate size of the basin. Thus, theoretically, for each of these graphs all points should lie on the fitted regression line. Values of stream

Fig. 16

Linear regression Log.N_o with O summary graph

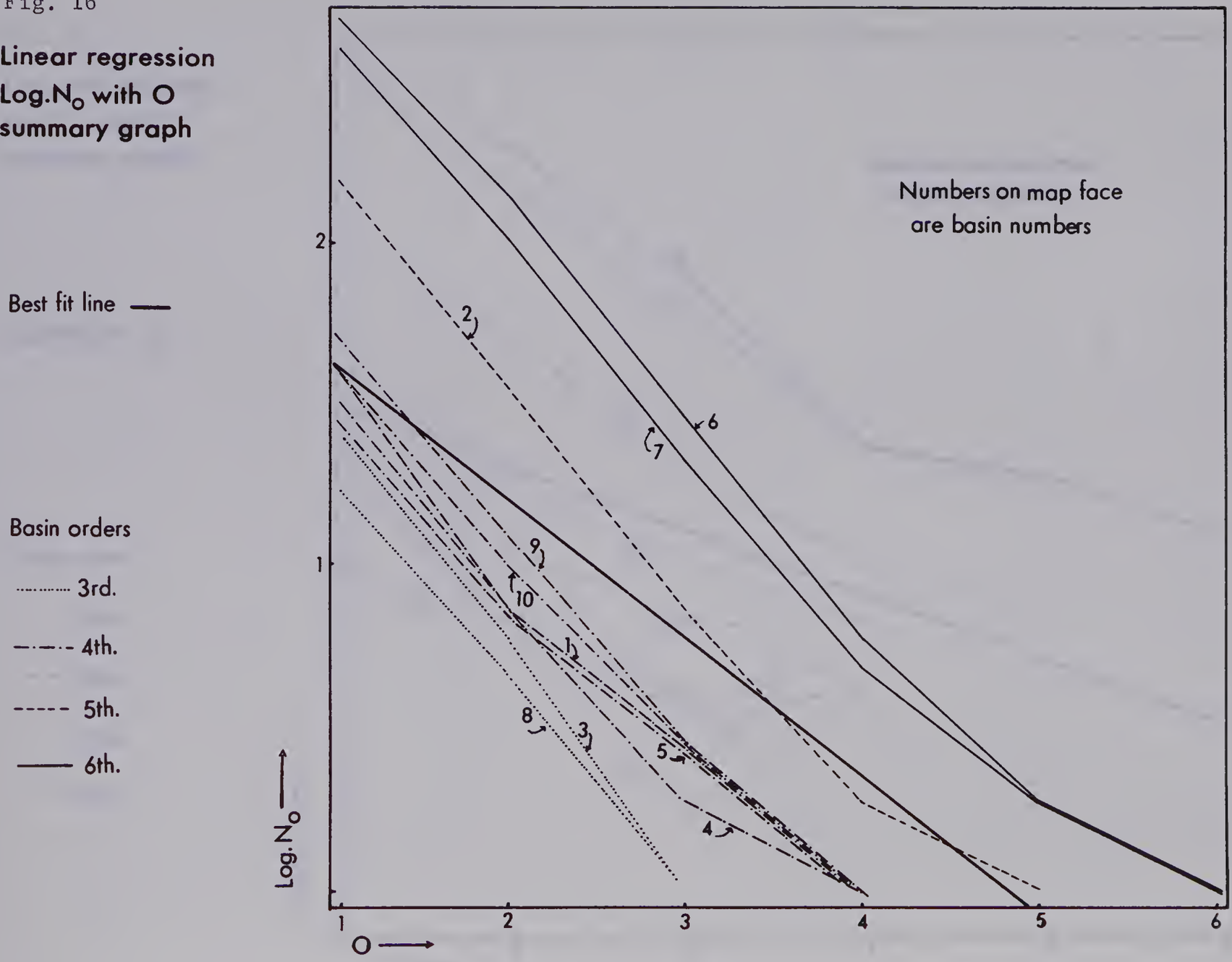
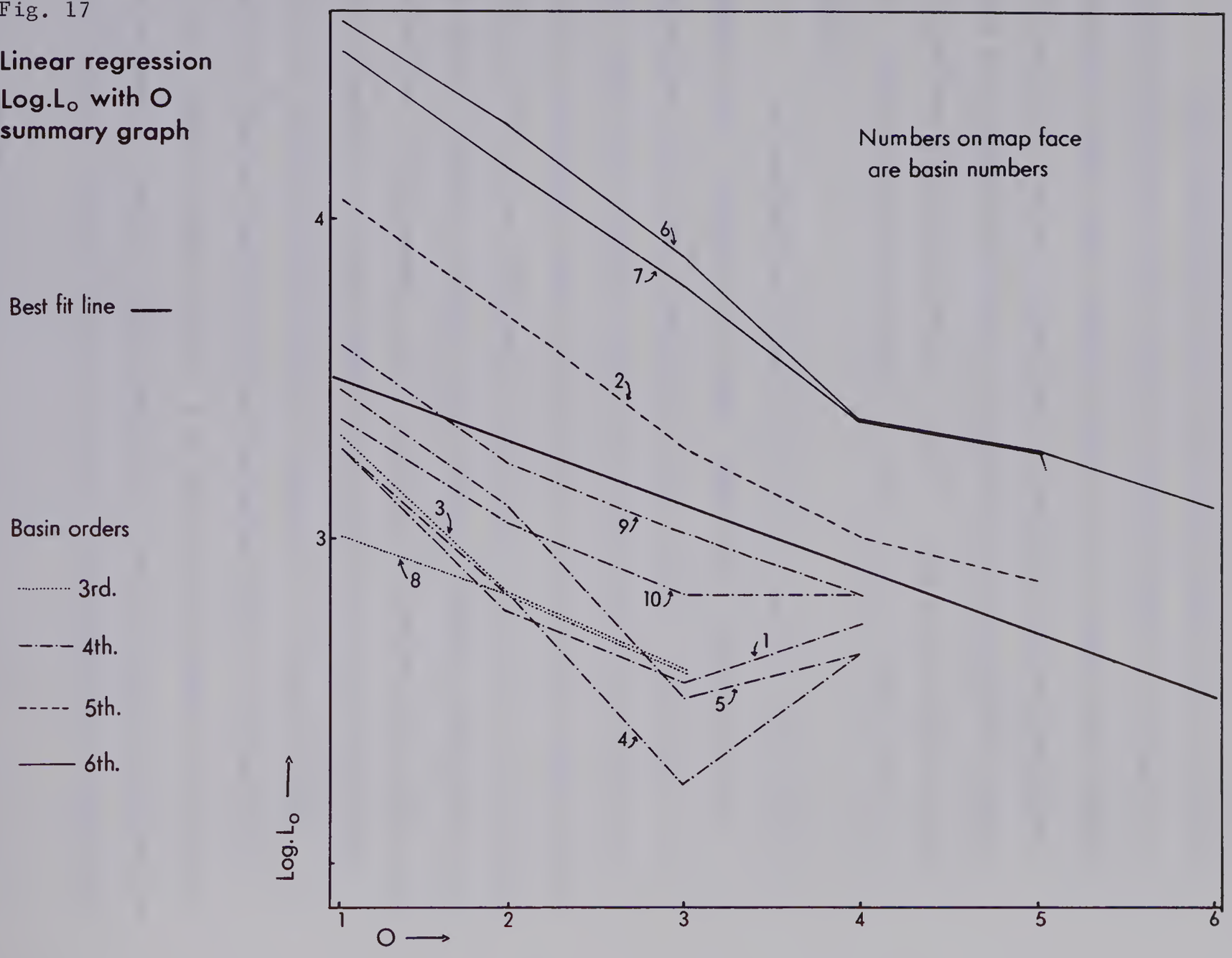




Fig. 17

Linear regression Log.L_o with O summary graph





length ratio and basin area ratio can be derived from the antilogarithm of the gradient of these fitted lines.

The divergences from the mean regression lines are most significant indicators of the regional geomorphological development. Although all basins lines parallel each other quite well, intercept the abscissa at s, and are in positions according to size as described, there is a marked up-concavity of all basin lines except for those of the smallest (3rd order) basins, 3 and 8. This means that the bifurcation ratios are generally higher for low orders and lowest for high orders[5].

An examination of aerial photographs and Fig. 15 can help to explain this. In most of the area, first order streams are over-represented due to the excessively rilled surface. These drain either into second order streams, which are correspondingly over-represented, or drain prematurely into a third, fourth, fifth or sixth order stream. Second order streams often enter immediately into fifth or sixth orders and do not coincide to form a third order as expected in Horton's random network. This anomaly, which occurs in all basins but 3 and 8, has been documented by Schumm (1956b), and agrees with the description of laterally expanding gullies given in Chapter I. It may, therefore, be suggested that up-concavity of the regression line is a characteristic of well dissected gully terrain, i.e. that the development of the drainage network in heavily eroding landscape is not topologically random, but (particularly where the prairie level has been reached) is a climatically

[5] Schumm (1956b) found similar trends in the badlands at Perth Amboy.

determined "finger-development". Basins 6 and 2 exhibit this characteristic quite noticeably, being the most active in laterally expanding.

An examination of Fig. 17 further documents this point. The fifth order basin (2) the sixth order basins (6,7)[6] and two of the fourth order basins (9 and 10) - in fact most of the "dominant" basins exhibit up-concavity. Of particular interest are the very short lengths of third order streams compared to the main channel in basins 1, 4 and 5, a trend that is again apparent in Fig. 18. Why these latter three diverge so far from the regressions may be explained in terms of the model presented in Chapter I and by examining Fig. 15. The growth of basins 4 and 5 in particular are being inhibited by the rapid extension of the first order tributaries of basin 6. Basin 1 is controlled in development by resistant lithology outcropping on the retreating scarp face A' - A" (Fig. 10) so that, despite the small freedom its headwaters have to expand into the upper prairie, its third order segment is shortened by a structural break caused by ss_1, ss_2 and ss_3 (see Fig. 3). Other basins apparently inhibited (3 and 8) do not exhibit the same characteristics as 4 and 5. However, basin 8 has compensated to some extent by elongating, and basin 3 by excessive development of tributaries downstream.

Fig. 19, average area drained by streams of a given order with order, gives the best fit to the regression for all basins, and only

[6] Basin 7 is somewhat of an anomaly in the stream length analyses as it is an empirically defined drainage area, not strictly a "basin", and hence the sixth order segment is severely truncated and is therefore far too short to fit the regression line. This illustrates the necessity for confining drainage net analyses to basins at their junction with a higher order stream or at some strictly defined temporary grade level (eg. entry onto alluvial fans).

Fig.18

Linear regression Log.$\bar{L}_o$ with O summary graph

Best fit line ——

Basin orders

..... 3rd.

—·—·· 4th.

----- 5th.

——— 6th.

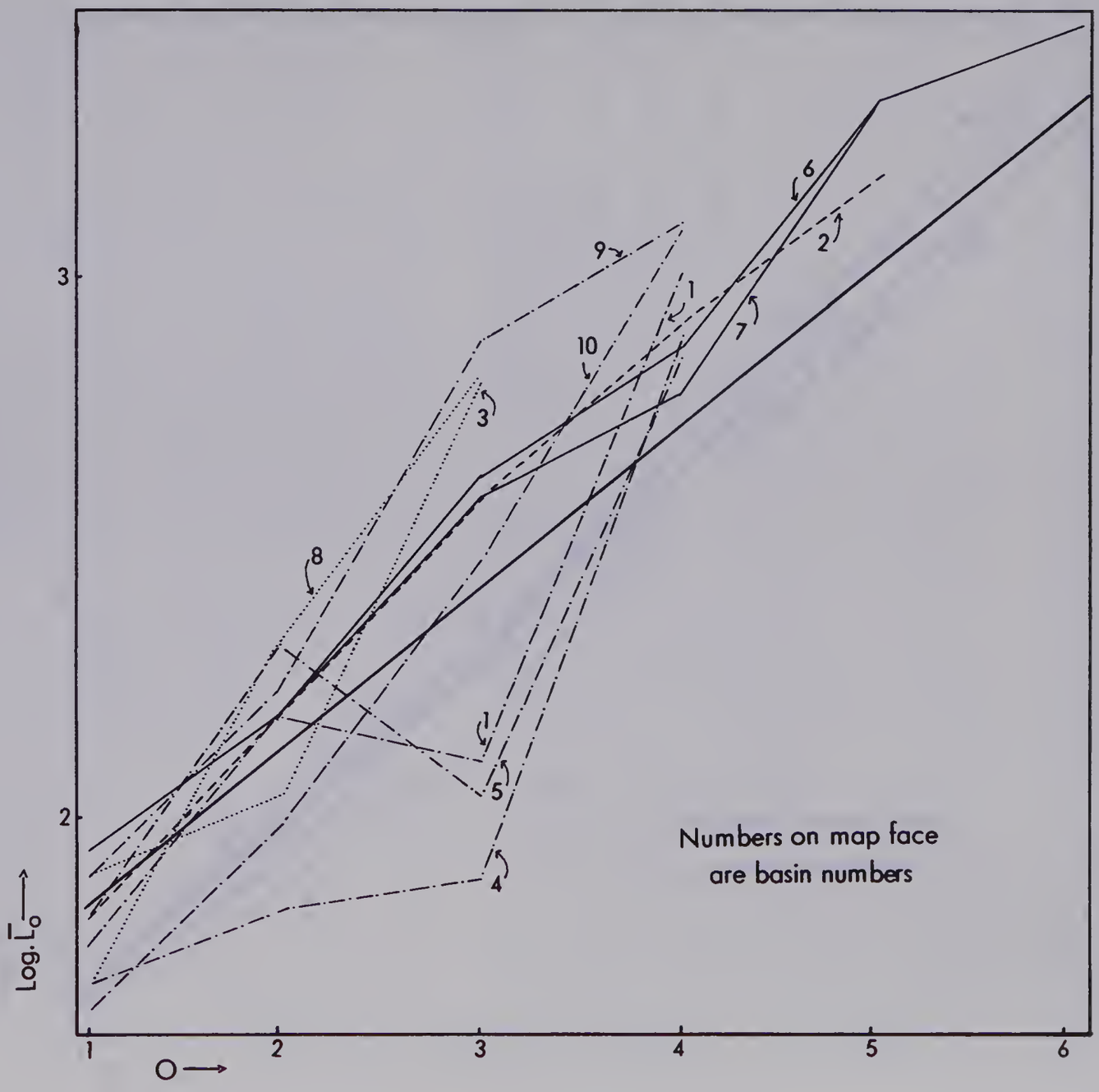
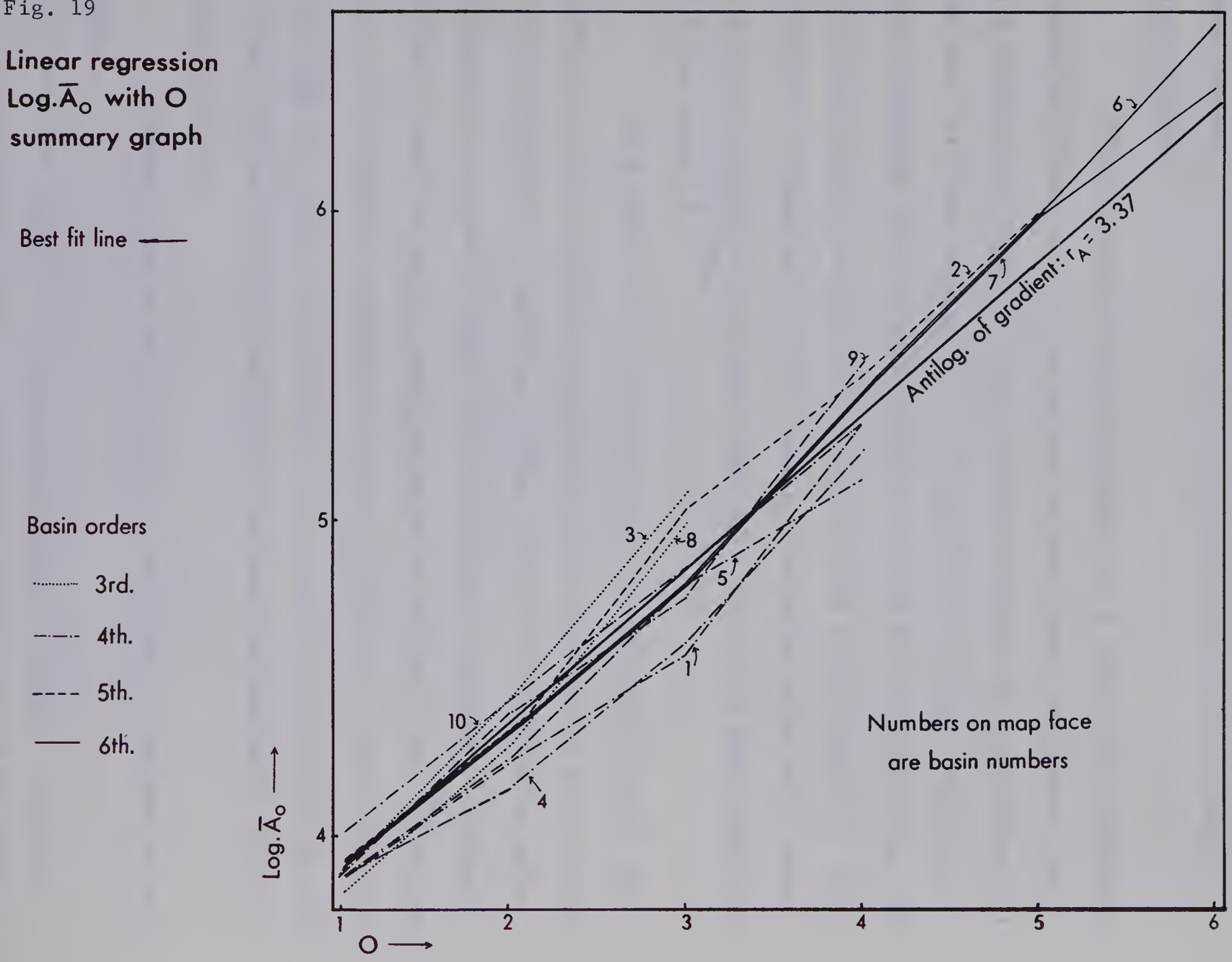




Fig. 19

Linear regression Log.$\bar{A}_o$ with O summary graph





slight evidence of up-concavity can be seen. The two third order basins (3 and 8) appear to be draining areas rather larger than the mean, whereas all but basin 9 of the fourth order basins are smaller than the mean.

The average morphometric ratios, r_b, r_L and r_A for each basin (Appendix A) tend to a certain extent to mask these divergences from the straight line since they are the antilogarithms of the gradients of the best fit lines through all the points for that basin. The ratios were determined for each basin and were plotted on histograms (Fig. 20). Bifurcation ratios ranged from 2.8 to 4.2, with a mean value 3.7. Similarly wide ranges existed for r_L and r_A, the modal value for r_L being 2.5 for all basins, and the mean value 2.2. For r_A the modal value was 3.4, the mean 3.7.

To a large extent the divergences in the ratios seen on the histograms can be explained by the few anomalies described above. However, it appeared on inspection of Fig. 1 that high bifurcation ratios occurred as expected in basins with the greatest elongation. Schumm's (1956b) elongation ratio r_e was considered an adequate index of this and was plotted with the morphometric ratios r_b, r_L and r_A (Fig. 21). Basins with high r_b, r_L and r_A were found to correlate quite well with the most elongate basins, and was considered an adequate explanation of the divergent values.

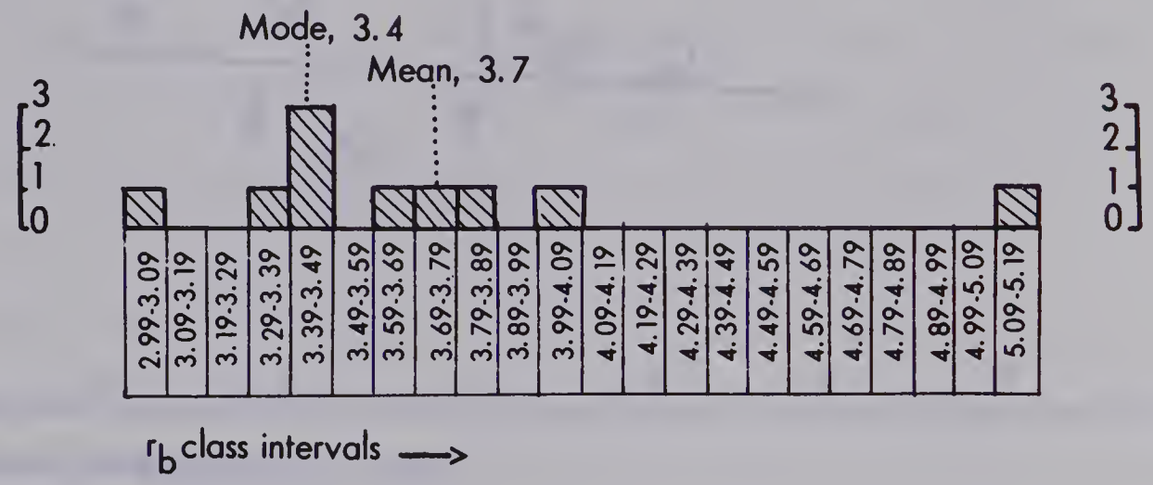
Several conclusions can thus be made from the regression analyses:

(1) The class intervals of the modified Horton (Strahler) ordering system are too large since the five basins demarcated order 4 are of quite different sizes. The characteristics of basins 4 and 5

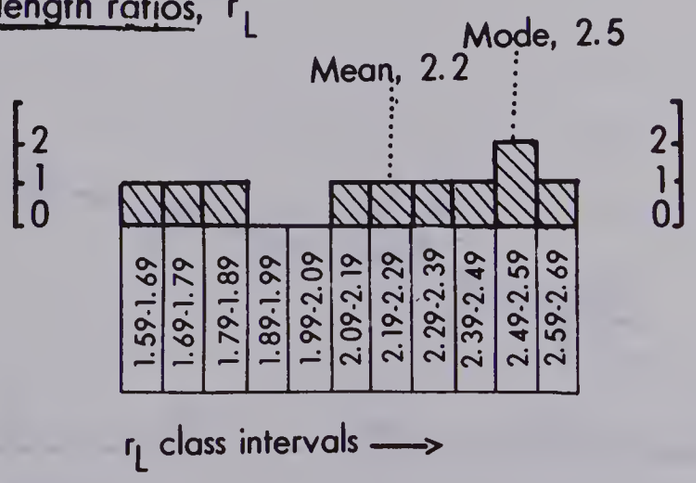
Fig. 20

FREQUENCY ANALYSIS OF MORPHOMETRIC RATIOS

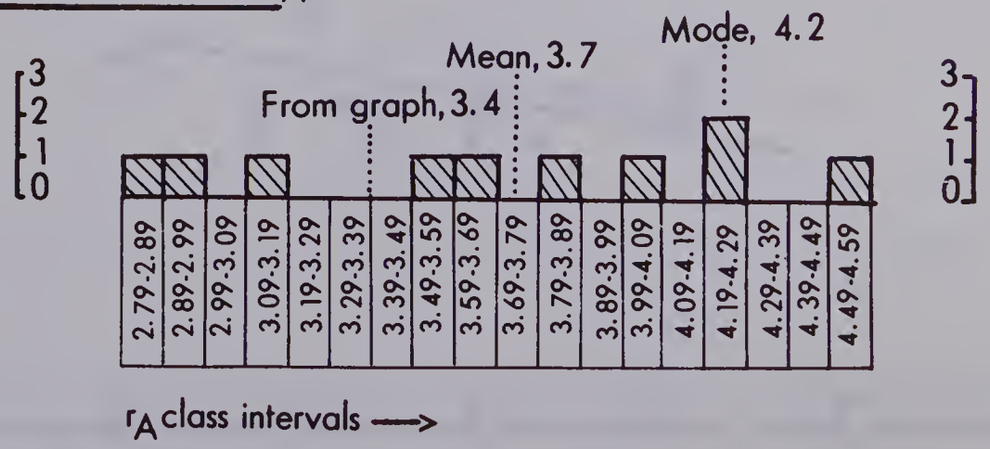
Bifurcation ratios, r_b



Stream length ratios, r_L



Basin area ratios, r_A







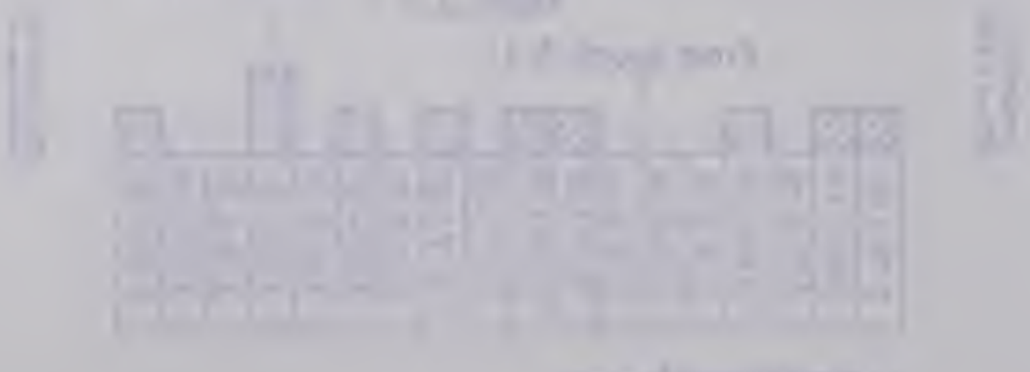
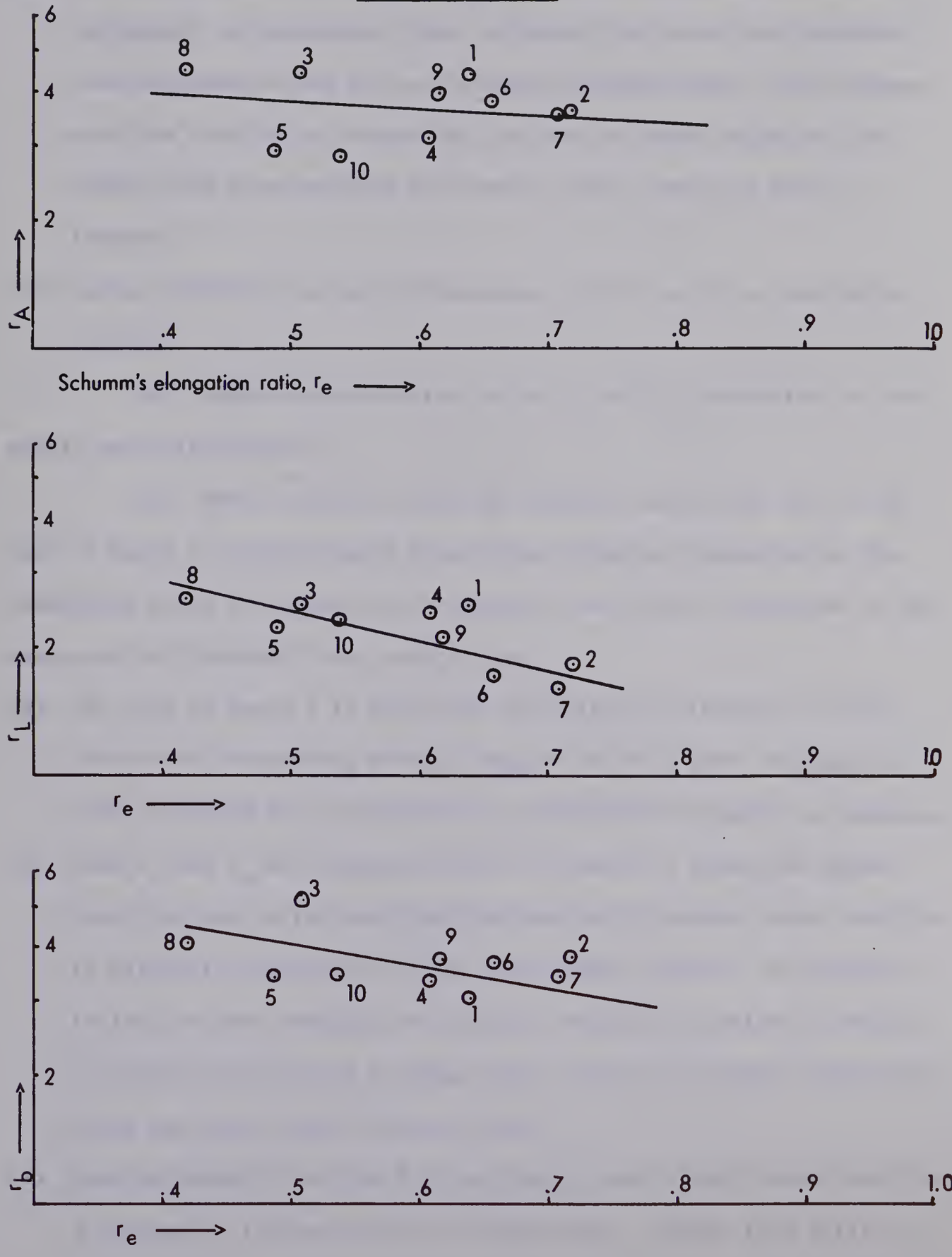


Fig. 21 VARIATIONS IN THE MORPHOMETRIC RATIOS WITH BASIN ELONGATION









are more akin to the third order basin 3, characteristics of basins 1, 9 and 10 being nearer to the fourth order norm for the area.

(2) The least restricted and dominant basins (2,6,9) have the better adjustment to regression lines, although low orders are generally over-represented and the main channel underdeveloped. This agrees with the conclusions concerning the mode of basin evolution for basins that have breached the prairie level that were made in Chapter I.

(3) Basins inhibited in their development (3,4,5 and 8) in particular exhibit:

(a) Under-representation in the L_o and N_o regressions in the middle order tributaries.

(b) Total area less than the regional mean (4,5) or, in the case of basin 3, overdeveloped first order tributary expansion in the downstream area; or in the case of basin 8, artificial elongation of the basin due to "pressure" from either side.

(4) The form of basin 1 is that most controlled by lithology on the theoretical retreating scarp. Despite the fact that the basin is free to expand into the prairie its third order segment is truncated.

(5) When L_o and $\bar{L}_o$ are regressed with 0 in basin 7, anomalies appear which are due to the empirical definition of drainage area resulting in artificial truncation of the sixth order segment. A corollary to this is that drainage net analysis should be confined to basins at their junction with a higher order stream or at some defined base level (eg. entry onto alluvial fans).

(6) Mean morphometric ratios for each basin tend to mask deviations from a regression line which may be significant. Ratios from first to second order compared to fith to sixth order, for example within

one area may be more geomorphically significant than the mean value.

(7) Variations in the average morphometric ratios for each basin may be considered a function of basin elongation, r_e.

(8) The graph of $\bar{A}_o$ with O gives the best fit to a regression line and exemplifies a consistency of areal division within the basins.

(9) Up-concavity of the regression lines is considered to be due to a combination of three factors;

(a) The natural tendency of modified Horton analysis to produce this in the higher orders.

(b) A sequential domination of the smaller basins by larger basins, limiting the growth of the latter - see Chapter I(C).

(c) A non-random network development ascribed to climatic control, where many first and second order streams flow prematurely into higher order streams, i.e. the typical "finger-gullies" of semiarid erosional landscapes.

(ii) Synthesis

In order to relate the consistency of the progressive decrease in lengths and areas with order within the basin (as described above) to the form of the channel at its emergence from the basin (Ch. III) it was necessary to establish in a clearer form relationships between basins. More specifically, relationships between A_s, the total area of each basin, and s, the order of its main channel, and between L_s, the length of the main channel and its order s were required, from which the basic relationship L_s to A_s follows.[7] Later, these two indicators

[7] The way this relationship is carried through the drainage net can be seen by examining the regression $\bar{L}_o$ with $\bar{A}_o$, which has the same exponent as L_s with A_s (0.4). See Appendix B.

are related to channel geometry variables (Ch. IV).

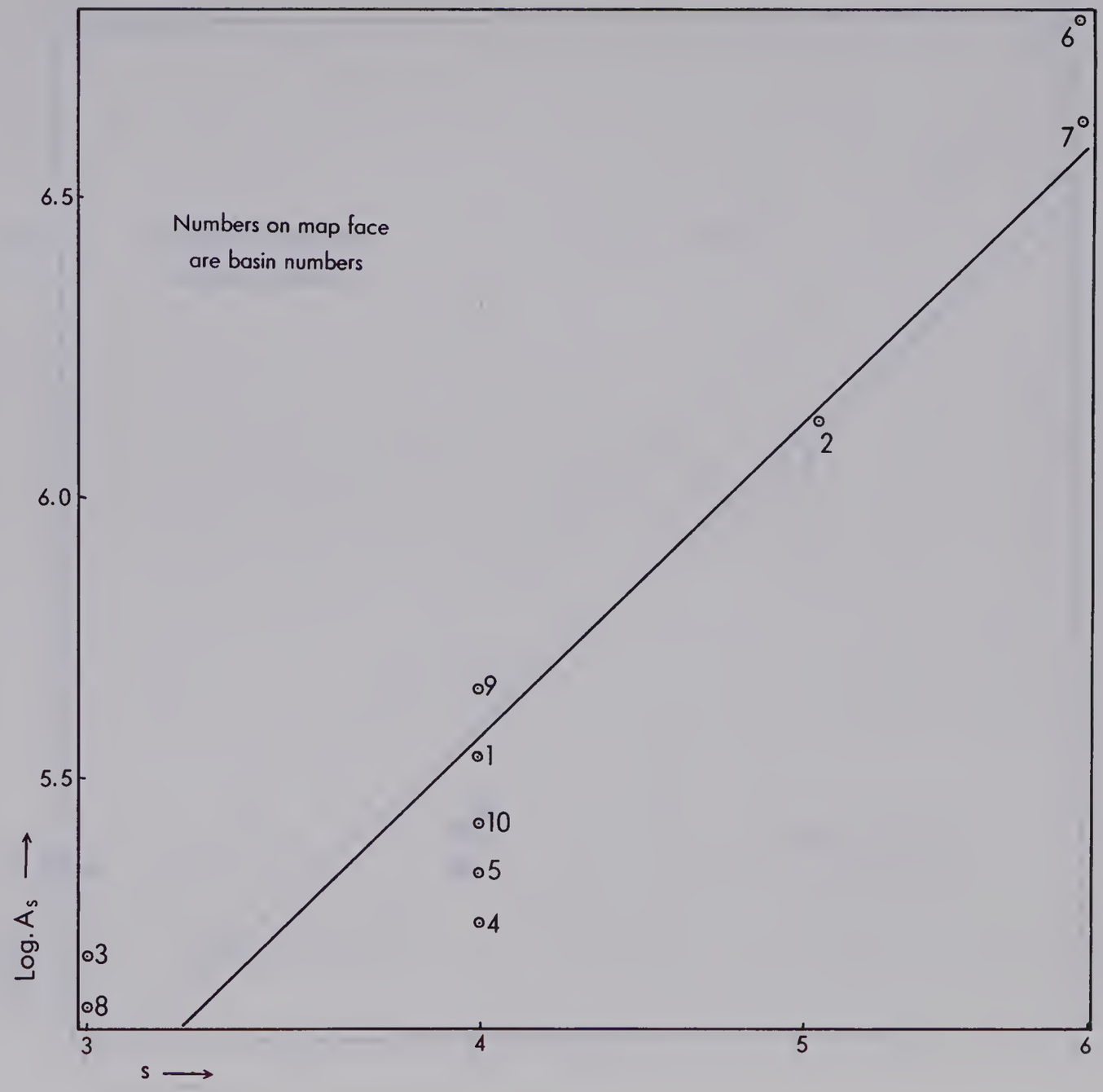
Despite the large rank intervals of Strahler's ordering scale and the consequent variations in size of basins with each order (s = 4 in particular) a good fit was obtained for regression lines A_s with s, and L_s with s (Figs. 22 and 23). This merited regression analysis of A_s with L_s which also correlates well (Fig. 24, see Ch. IV).

As a result, despite different levels of interbasin reaction and mode of development of basins in the study area, certain consistencies in the planimetric form of the drainage nets have been established.

Fig. 22

Linear regression

Log. A_s with s



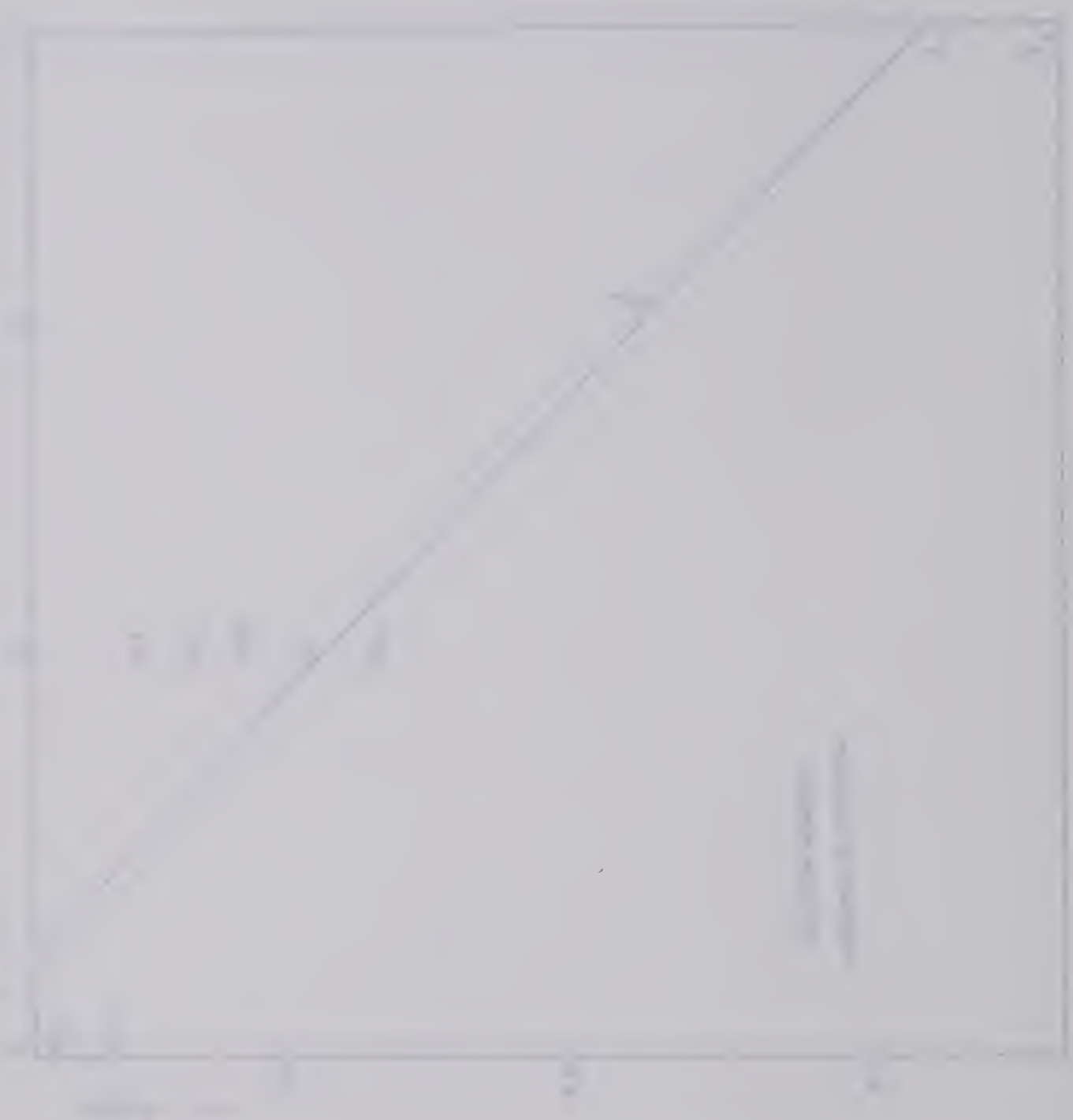


Fig. 23
Linear regression
Log. L_s with s

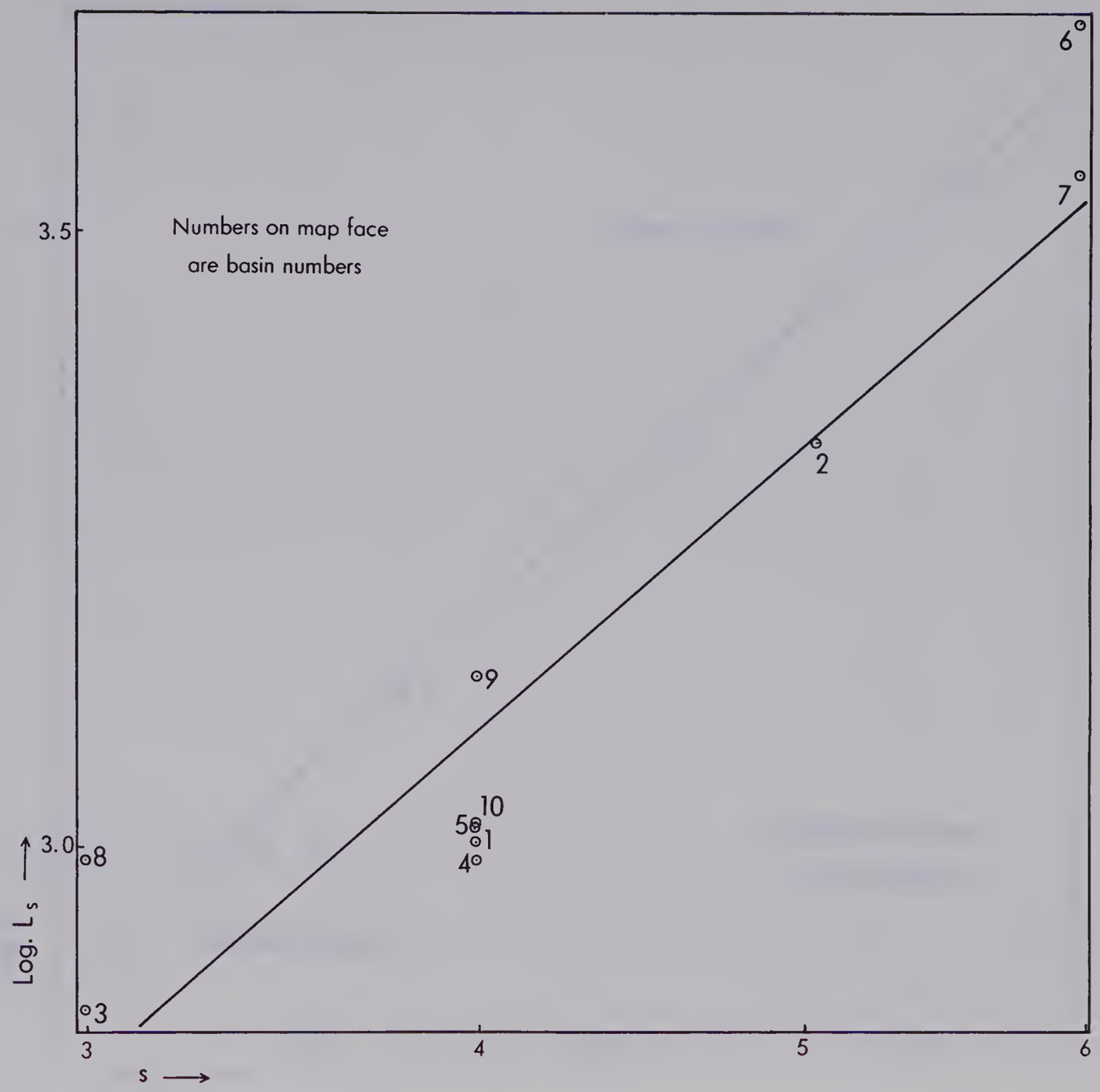
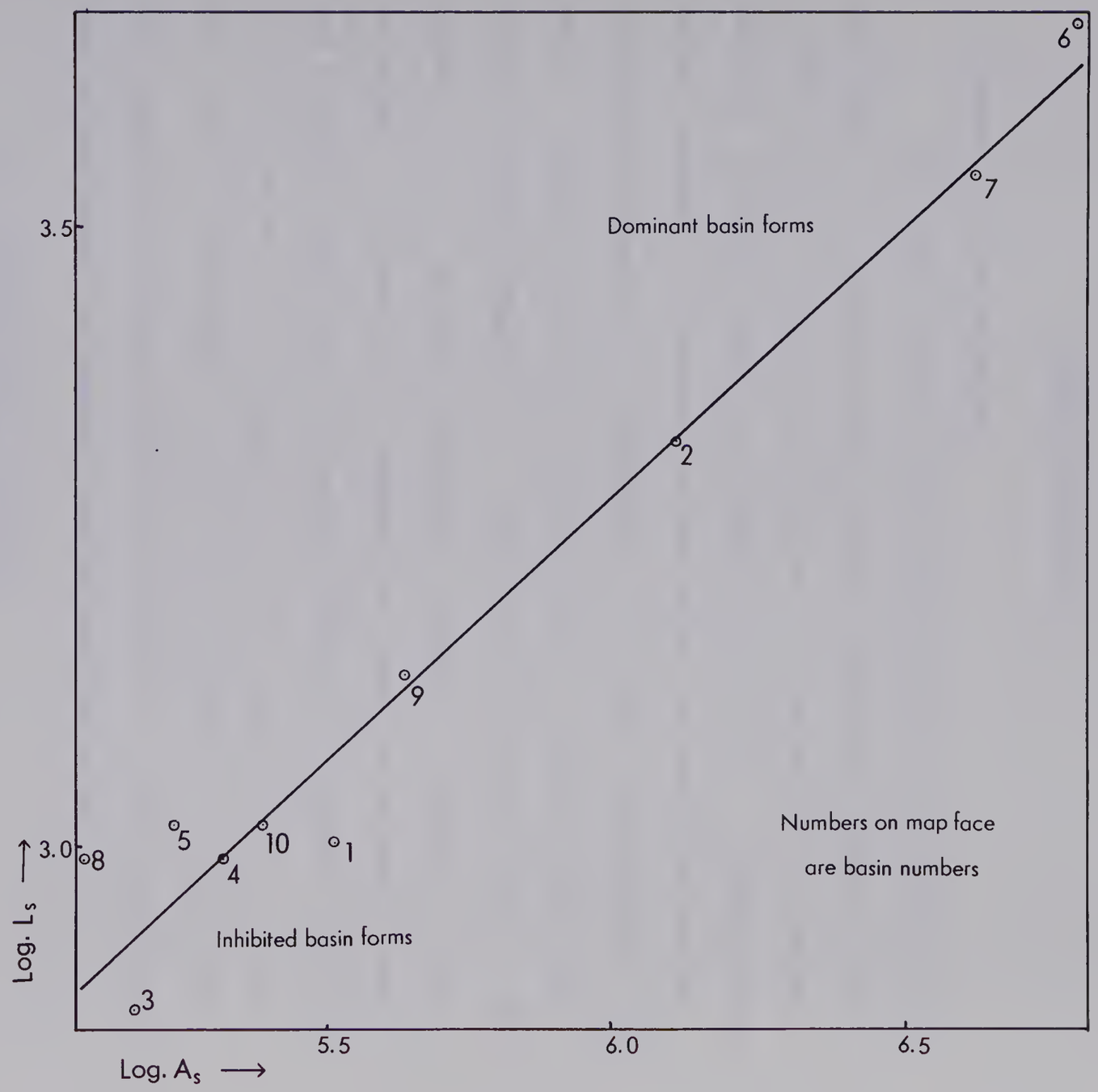




Fig. 24

Linear regression

Log. L_s with Log. A_s



Relationship between plotted variables:

$$L_s = 5.0(A_s)^{0.4}$$

significant at 0.1% level



CHAPTER III

CHANNEL MORPHOLOGY

A. Theoretical Considerations

The shape and size of the cross-section of any channel is an adjustment to the volume of the flow, the quantity and nature of the sediment passing through the section, and the composition of the bank material, including vegetation (Leopold et al, 1964). In simple stable natural channels with moveable bed and banks, two conditions must be satisfied simultaneously; transmission of the flow, and stability of the banks. Channels that ideally satisfy these two conditions have been called "threshold channels"; i.e. those within which all points on the perimeter are at an erosion threshold, although no erosion is in fact occurring. In actuality this hypothetical condition is rarely found, and a natural channel not only carries sediment but migrates laterally by erosion of one bank and deposition on the opposite bank. In this case, the "no bank erosion" condition is replaced by a continuous dynamic equilibrium between erosion and deposition. Where this occurs the form of the channel is constant, but its position is not.

(i) Channel-forming Discharge

During dynamic equilibrium, the form and size of the channel cross-section represents an adjustment to some dominant flow condition of the channel. Confusion arises, however, in the definition of this. For example, Leopold et al, (1964, p. 296) state that "...channel forms," include "...meander wavelength [which] is empirically related to the square root of the effective or dominant discharge". Dominant discharge

has been defined by Inglis (1941, p. 110; cited in Carlston, 1965) as "...slightly in excess of bankfull discharge," but Leopold et al, (1964, p. 241) consider the "...effective discharge can be approximated by bankfull discharge". Carlston (1965) however, suggests that both dominant and effective discharge can probably be safely approximated to bankfull discharge. Dury (1968) has examined the concept of "bankfull" discharge and its effect of channel formation in some detail. He observed that "...conditions at bankfull are a species of norm about which relevant values fluctuate", and in a later publication has concluded that "...it is at bankfull that water is in contact with, and is acting upon, the whole perimeter of the channel." (Dury, 1969, p. 421)

The bankfull stage is extremely significant geomorphologically since it determines the ultimate channel form and size, and when regressing discharge with other channel variables in a downstream direction, or with basin variables, it becomes necessary to ensure that the discharge value used is always bankfull (Q_b).

Manning's equation (see Leopold, et al, 1964, p. 158);

$$Q = A_c \times \frac{1.486}{n} \cdot S^{1/2} \cdot R^{2/3} \qquad (1)$$

has frequently been used to assess bankfull ("channel-forming") discharge on streams where direct velocity measurements are not possible. It was necessary to adopt this equation to compute Q_b through each cross-section on each of the dry channels in the study area due to the fact that the bankfull stage was not attained during the field season.

The technique involves the determination of Manning's largely empirical roughness coefficient, n, at each cross-section studied. Barnes (1967, p. 1) stated that,

> "...at the present state of knowledge, the selection of roughness coefficients for natural channels remains chiefly an art. There are no resistance diagrams, or quantitative relationships available ... consequently the ability to evaluate roughness coefficients for natural channels representing a wide range of conditions must be developed through experience".

A useful guide in choice of n was found in the table produced in Te Chow (1964; see Appendix C) and in the series of photographs presented in Barnes (1967) both of which were studied for purposes of assigning n values in the field as consistently as possible.

The Manning equation (1) also involves measurement of S, the slope of the energy grade line through the channel cross-section. This has been assumed by several workers (e.g. Leopold and Maddock, 1953, p. 14) to approximate the slope of the water surface, which can be determined by measuring the slope of the bank adjacent to the channel by means of an Abney level. The hydraulic radius of the channel (R) is computed by dividing the cross-sectional area of the channel (A_c) by the length of the wetted perimeter (P), both of which can be derived by plotting the channel cross-sections to scale on graph paper.

Thus a practical index of the bankfull discharge is afforded by the Manning equation, and these computed values can be compared with the form of the channel, its pattern, and (in Chapter IV) with drainage basin characteristics.

(ii) Channel Form

It was explained above that the form of the channel cross-section reflects the bankfull discharge. The nature of the bank material, however, influences the overall shape to which the cross-section will tend, for example, channels with cohesive silty banks and sand beds will be narrower in general than comparable ones with sand banks and sand

beds (Schumm, 1960). Since each channel in the study area has similar bed and bank material, then the form ratio (w/d) might be expected to be similar for the straight reaches of each stream. However, the rate of increase of width, depth and velocity with bankfull discharge in a downstream direction has been studied in great detail by Leopold et al (1964, Ch. 7, p. 202), and it is apparent that since the width of a channel increases faster than depth in a downstream direction "...the appearance of rectangularity [of the cross-section] increases somewhat as the river gets larger downstream..." i.e. the form ratio will increase for greater length of main channel. The ten channel segments in the study area were examined for evidence of such a trend.

The general form characteristics of the channel are governed by bankfull discharge, composition of the bed and banks, and the location of the section on the main channel; smaller features may develop in the channel that are not necessarily formed at bankfull discharge, and are often independent of the composition of the bed and banks. For example, a straight or non-meandering channel characteristically has an undulating bed, and alternates along its length between deeps and shallows, spaced at a repeated distance equal to 5 or 7 widths (Leopold, et al, p. 207). This phenomenon also occurs in meandering channels, but this should be expected since in these cases pools are generally associated with bends, where there is an obvious tendency to erode the concave bank. The pool and riffle sequence at 5 to 7 widths has been well documented in perennial streams, but is not particularly evident in ephemeral channels viewed in their usual dry state. Leopold et al (1964, p. 209) suggest, however, that an analogous feature occurs; thin surface accumulations of coarse material in the form of gravel bars have been found

that similarly alternate with pools at 5w to 7w. These features "...remain with minor change from year to year". The gravel riffles were regarded as "...a kind of kinematic wave in the traffic of clastic debris...a wave that travels downstream faster than the movement of the particles themselves." (p. 212). This theory is in agreement with conclusions made by Leopold and Miller (1961) concerning wave phenomena in the rising flood stage along the ephemeral channels of the Arroyo de los Frijoles and the Rio Galisteo near Santa Fe, New Mexico. In these areas, channel bank caving and small slumps were noted frequently as well as the gravel riffles, and this was accredited to both the nature of the kinematic waves and the inability of the sparse vegetation to consolidate channel banks.

The effect of vegetation is quite important in ephemeral channel formation, since the poorly consolidated banks of channels where little vegetation is growing implies an easier width adjustment in these channels than those of more humid climates. Where banks are held together by roots, an increase in discharge is more likely to be associated with simultaneous depth and width adjustments. Channel adjustments to varying discharge such as these were sought in the study area.

(iii) Channel Pattern

The development of the channel pattern (i.e. whether meandering, braided or straight) represents an additional mechanism of channel adjustment which is related to channel gradient and cross-section.

> "The pattern itself affects the resistence to flow, and the existence of one or another pattern is closely related to the amount and character of the available sediment and to the quantity and variability of the discharge." (Leopold, et al, 1964, p. 281)

Straight channels occur rarely in nature, being essentially unstable, and in general the thalweg will wander from one bank to another, and opposite the deepest points bars or accumulations of mud develop which alternate downstream. This initiates a pool and bar sequence which in turn may lead to the development of meanders. The typical heterogeneous grain size associated with pool and riffle sequences implies non-cohesive beds and banks. Moveable beds and banks are characterised by high width to depth (w/d) ratios "...this in turn tends to promote a braided pattern". (Leopold et al, 1964, p. 282). The gradation from straight to sinuous, from sinuous to meandering, and from meandering to braided occurs gradually and it is difficult to pinpoint cause-and-effect relationships. Factors other than high form ratios encourage braiding; for instance, a change from dense to sparse vegetation along the banks, or a sudden large change in flow resistance, such as islands, will encourage channel division "...but again, the unravelling of cause from effect is usually difficult or impossible." (Leopold et al, p. 282).

In 1952 Rubey suggested that braided channels may indicate equilibrium channel conditions rather than aggradation. In a later study Leopold and Wolman (1957) described in detail processes of braiding caused by bank caving and bar shifting. They found that the slope of the braided reaches studied was considerably greater than that of the undivided reach with corresponding degradation and aggradation below and above the divided segment. This agreed with Rubey's (1952, p. 126) conclusion that "...on the average the slope opposite islands is steeper by something like 5 to 10 per cent." In addition, all examples showed

that channel division was associated with an increased width of water surface, and increased variable depth, and as a result an increase in the size of the cross-section. As Leopold and Wolman (1957, p. 295) conclude,

> "...changes in nearly all the hydraulic parameters occur when a channel divides", and this is, "a type of adjustment that may be made in a channel possessing a particular bank material in response to a debris load too large to be carried by a single channel. Braiding then represents a response or adjustment among the controlling variables which may provide an equilibrium condition over a period of time."

Another type of adjustment made by the mature channel is the development of meanders. Because most streams are sinuous to some extent, the term "meander" is reserved for more symmetrical curves, in which case the meander wavelength in particular is closely related to both bed width and discharge (Leopold, et al, 1964; Leopold and Wolman, 1957; Dury, 1967; Carlston, 1965). The formation of meanders has been accredited to helical flow (Leviavsky, 1955) although it is not clear how this can cause the observed wavelength and width relationships. Forces determined by velocity distribution, including helical circulation, can account for the shape of a cross-section in a meander, the deposition and erosion pattern, and the downvalley migration, but again these alone are not sufficient to explain the dimensions and proportions of meander forms. Considerations of energy conservation may prove useful in the future to explain meander relationships.

All aspects of channel pattern, straight, braided and meandering were exhibited on the ten stream segments studied in the Steveville badlands. Computed velocity and discharge values, and the associated slope and channel roughness characteristics, may throw further

light on the variations in both channel form and pattern observed along the channel segment. For this purpose the channel segments were mapped in detail in the field.

B. Field Measurements and Channel Mapping

The point where channel and watershed intersect was located at the mouth of each study basin. At this point a reference peg C was placed on one bank of the channel, and a corresponding peg H placed so that CH was 25 feet and paralleled the fenceline along the bottom of the scarp face. The distance from C to the fenceline was noted. A reference grid of 25 foot by 25 foot squares was constructed in the field, using a tape and Brunton compass, which was based on CH and which in all cases except segments 2 and 4 (where the grid turned out to be at a slight angle to the fence) paralleled the fenceline. In some cases the fence cut across the channel in the study segment, in other cases it was further south, but the grid was in reference to the fence[1] because this could be easily identified on the map and aerial photographs. The grid was constructed to such a size as to include at least 50 feet of channel above and below the watershed intersection, so that pegs A,B,I and J were usually within the basin, CH on the watershed, and pegs D,E,F and G outside the basin. Thus a series of cross-sections would be obtained that represented the channel at the watershed intersection.

The first cross-section was taken along the line AJ, and thence at a consistent distance along the channel floor - 15, 12 or 10 feet

[1] Except in the case of sub-basin 7 where the fence did not occur near the channel segment.

depending on the sinuosity of the channel (10 cross-sections were required that would fall within the 25 foot x 100 foot grid). The point of intersection of the line AJ with the channel was located, and the first cross-section was measured perpendicular to the thalweg at this point. A peg was placed on each bank at "bankfull". This was the line above which vegetation occurred with little or no evidence of floodwash around roots, and below which vegetation was largely absent (this part was usually characterised by an increase in slope). The level of bankfull was quite difficult to pinpoint exactly in several cases, but usually where this occurred levelling from the peg on the other bank could approximate its location. Once the pegs had been placed, a wooden rod marked at 6 inch intervals was placed between the pegs and levelled using the bubble on the Brunton compass. The depth of the channel at each 6 inch interval was determined by measuring from the rod to the channel bed, and readings were accurate to the nearest half inch. When one cross-section had been measured, the chosen distance (15, 12 or 10 feet) was measured along the thalweg and this point became the location of the next cross-section. In all channels except segment 6, ten cross-sections were measured in this manner. Pegs on either bank were fitted into the grid by measuring the distance from the two nearest reference grid points to each cross-section peg.

The form of the channel was noted in detail as the cross-sections were measured. Field sketches of bank form,pool and riffle sequence, vegetation, and obstacles to channel flow were made for each channel segment, and the thalweg followed from one section to the next.

Using an Abney level the slope of the bankfull stage through each cross-section was measured. This was facilitated by placing a pole

at eye level at a cross-section peg, and then measuring the slope to it from the next cross-section peg, and so on. The mean of the slopes above and below a peg represented the slope of the bankfull water surface through the cross-section. Slopes 100 feet above and below the study segment were also measured. Using Te Chow's table (Appendix C) for assessment of the channel roughness coefficient, n, each cross-section was accorded a value of n taking into account conditions above, below, and through the section.

Channel segment 6 was too large to fit in the 25 foot by 100 foot grid that was used in the other channels, and consequently a grid was constructed 250 feet by 100 feet with pegs at 50 foot intervals. A terrace-like feature was evident in this and channel segment 7, so that this level was used as the reference level for the cross-sections, and the bankfull stage noted on the bed within the section. The cross-sections were still measured perpendicular to the present channel thalweg, but slope S was not measured at peg locations (on the terrace) but between cross-sections at the bankfull location as described above. Thus the form of the terrace could be mapped, and bankfull cross-sectional area and water surface slope for the present channel still evaluated within this. Fifteen cross-sections were measured on channel segment 6 at 30 foot intervals.

Once all the data had been obtained, the channel segments could be mapped in some detail. By marking out the reference grid on graph paper, distances that had been taped from the reference grid pegs to the channel cross-section pegs could be drawn in as an arc using a pair of compasses. The intercepts of these arcs represented the location of the cross-section pegs. Between the cross-sections the details of channel form that were noted in the field could be drawn in. At a

different scale the cross-sections were constructed on the graph paper and the area of the cross-section determined by counting squares and multiplying by a scale factor. The hydraulic radius (R) of the channel was evaluated for each section by measuring the wetted perimeter (P) and dividing this into the cross-sectional area, A_c. For each cross-section on each segment V_{bc} and Q_{bc} were determined using the Manning equation, and were compared to variations in channel form and habit. Mean values of Q_{bc} and A_c were determined for each channel segment, and in Chapter IV these are related to A_s, L_s and s.

C. Discussion of the Channel Forms

All the field data are summarised on Figs. 25 to 45. Each channel segment will be described in detail and channel form characteristics related to the conclusions made in section A.

Channel Segment 1 (Figs. 25 and 26)

The moderately sinuous (1.2) channel is incised into the alluvium at the mouth of basin 1, its deep entrenchment being due to a rapid increase in slope immediately above the study segment (caused by the indurated sandstone bands ss_1 ss_2 and ss_3) and a high bankfull velocity results; calculated average was about 4 ft./sec. for each cross-section. Channel slope, S, is high, the steepest of all the channels, - and varies from 0.05 to 0.06 ft./ml. through the sections. Channel cross-sectional form is correspondingly near-parabolic, with values of hydraulic radius R close to unity (0.51 to 0.84), and with generally low form ratios (varying from 8.3 to 29.2). The channel thus appears quite efficient, but on assessing n, the moderately high sinuosity and rough banks and bed resulted in high n values. Slope

Fig. 25 **CHANNEL SEGMENT 1**

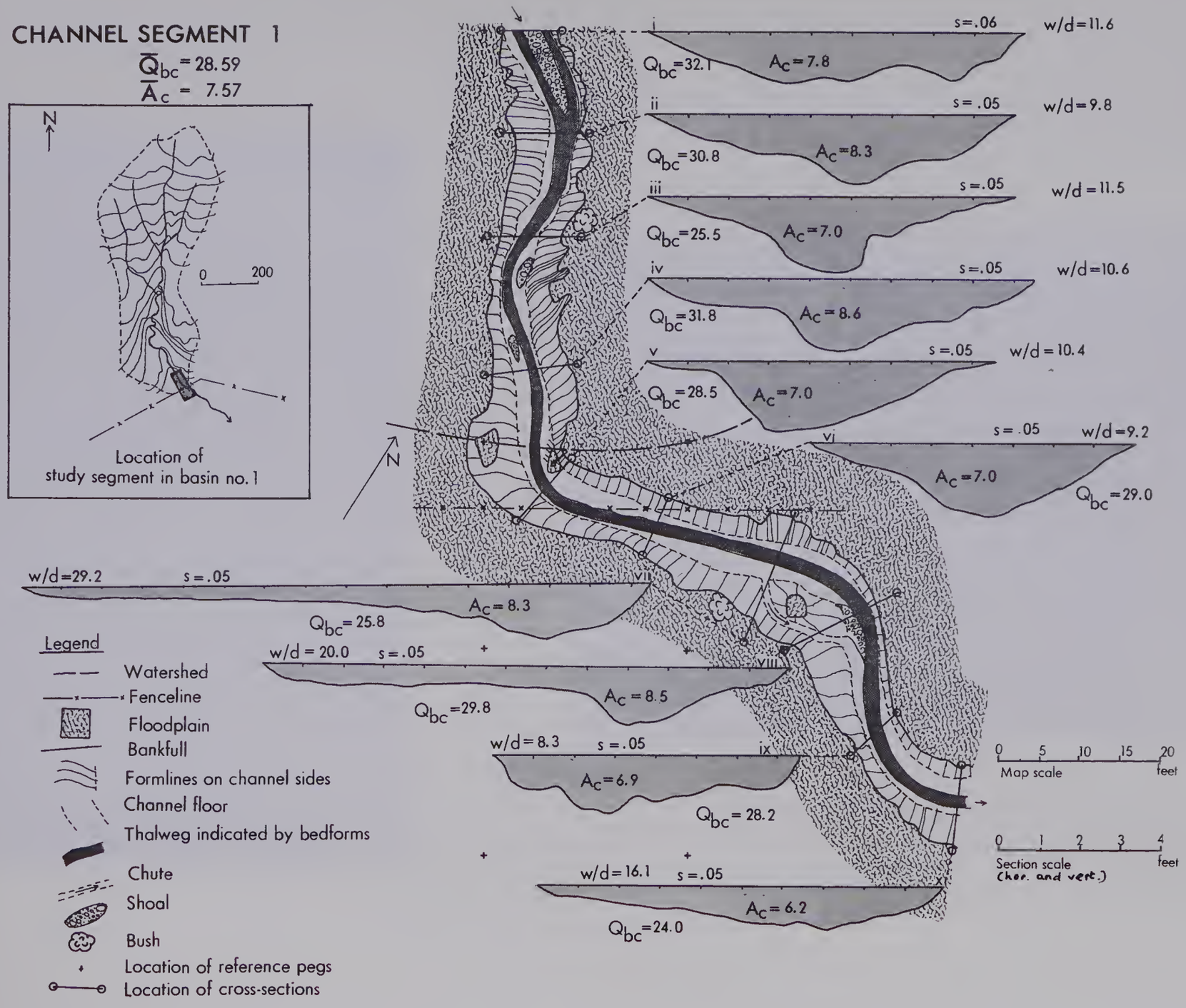




Fig. 26a Basin 1, showing channel at its watershed emergence (near fenceline).



Fig. 26b Channel 1 at the fenceline, facing downstream.

through the cross-sections was greatest at (i) and (v), and least at (iii) and (x), and computed velocities were correspondingly high and low at these locations. There is no obvious reason for these slope differences, although (i) is a riffle location and should therefore be expected to have a high S and large cross-sectional area (Rubey, 1952), which it does. Section (iii) is located at a bend and has a low A_c and computed velocity was lower here, again agreeing with Rubey (1952) and Leopold and Wolman's (1957) discussions of pool, riffle and braiding characteristics on sinuous channels. No consistent pool and riffle sequence was seen in this channel.

Computed velocity increased at the first channel bend, but was low for (vi) and (vii), where a chute has developed and the main channel seems to have aggraded due to flow diversion through the chute. Thus slopes are not high at this bend, but cross-sectional area is artificially large (8.3 and 8.5 $ft.^2$). Development of a chute might be thus expected to cause readjustment of channel variables in the same manner as braiding.

Computed discharge at bankfull varied from 24.0 to 30.8 $ft.^3$/sec. These values are extremely high considering the size of the drainage basin (see Ch. IV) and the channel segment should be considered anomalous due to structural control just above the study section (Fig. 26a).

Channel Segment 2 (Figs. 27 and 28)

The sinuous (1.3) channel segment at the mouth of basin 2 is similarly well-defined and almost parabolic, and form ratios are very low (5.8 to 15.5). Banks are heavily gullied as in channel 1, and the nature of these peripheral gullies is such that a 'ledge' appears

Fig. 27 **CHANNEL SEGMENT 2**

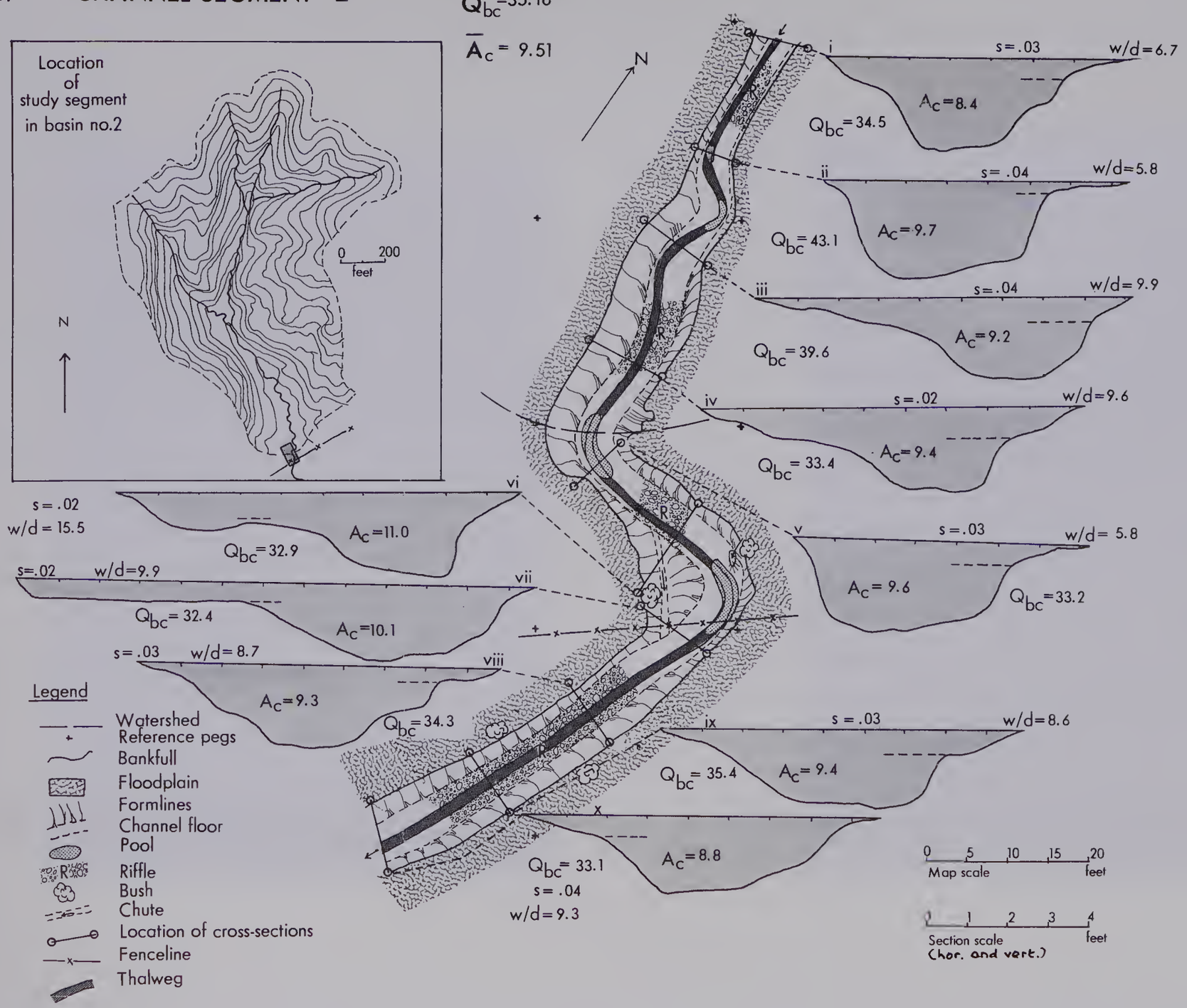




Fig. 28a Basin 2, showing widening of valley downstream as channel emerges from the watershed.



Fig. 28b Cross-section of channel 2 at the watershed intersection. Note peripheral gullies, facing upstream.

just below bankfull which could be interpreted as a terrace of some kind if field observation had not revealed otherwise (dotted line on sections, see Fig. 28b). Banks are moderately cohesive, but a slump block has developed at the third bend similar to that at the third bend in Channel 1. These blocks cause local eddying and may alter the thalweg location.

Associated with the high sinuosity a pool and riffle sequence is apparent, the recurrence interval being 5.5w. The gravel bars are similar to those described by Leopold et al (1964), being merely superficial features largely composed of sandstone and clay ironstone fragments, and may be indicative of kinematic wave conditions at rising stage. Associated with the high sinuosity and rough nature of the bed, n values are high. Slopes through the cross-sections are generally low, being a minimum at (iv) (v) (vi) and (vii). Cross-sections (iii) and (vi) occur below a riffle and represent decrease in slope adjacent to the pool at the first bend. Cross-sections (vi) and (vii) are similarly located where in this case a chute has developed in the same way as in channel 1, and hydraulic variables experience the same changes here (velocity decrease with cross-sectional area increase).

Computed bankfull discharge ranges from 32.4 to 43.1 $ft.^3$. sec. throughout the stretch, being highest on straight riffle stretches and lowest at bends. Mean Q_{bc} is 35.2 $ft.^3$/sec. compared to 28.6 $ft.^3$/ sec. for basin 1.

Channel Segment 3 (Figs. 29 and 30)

The channel draining basin 3 is wide, ill-defined, and shallow, with form ratios being large (13.1 to 32.0) and hydraulic radius low

Fig. 29

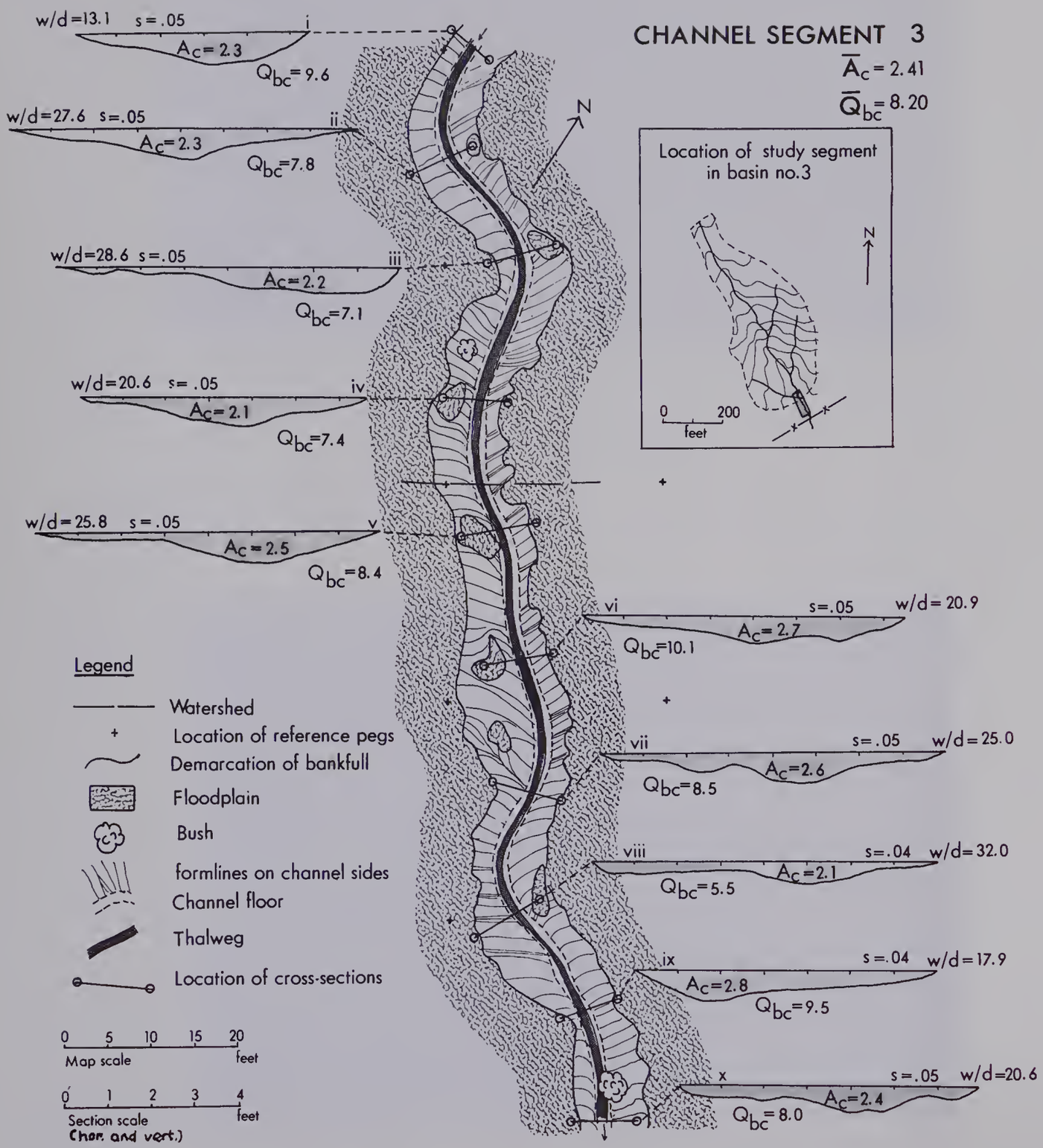

CHANNEL SEGMENT 3
$\bar{A}_c$ = 2.41
$\bar{Q}_{bc}$ = 8.20
Location of study segment in basin no.3
N
0 200 feet
w/d=13.1 s=.05 i
Ac=2.3
Qbc=9.6
w/d=27.6 s=.05 ii
Ac=2.3
Qbc=7.8
w/d=28.6 s=.05 iii
Ac=2.2
Qbc=7.1
w/d=20.6 s=.05 iv
Ac=2.1
Qbc=7.4
w/d=25.8 s=.05 v
Ac=2.5
Qbc=8.4
vi s=.05 w/d=20.9
Ac=2.7
Qbc=10.1
vii s=.05 w/d=25.0
Ac=2.6
Qbc=8.5
viii s=.04 w/d=32.0
Ac=2.1
Qbc=5.5
ix s=.04 w/d=17.9
Ac=2.8
Qbc=9.5
x s=.05 w/d=20.6
Ac=2.4
Qbc=8.0
Legend
Watershed
Location of reference pegs
Demarcation of bankfull
Floodplain
Bush
formlines on channel sides
Channel floor
Thalweg
Location of cross-sections
0 5 10 15 20 feet
Map scale
0 1 2 3 4 feet
Section scale
(hor. and vert.)



Fig. 30a Basin 3, showing point of emergence of channel from the watershed.



Fig. 30b Channel at (vi) (book). "Bankfull" was difficult to locate on this channel.

(0.25 to 0.4). Thus the bankfull position was quite difficult to locate exactly in several cases. The channel is almost straight (1.1) and has little rock debris in its course. Thus the allocated n values are quite low, and since the gradient S of the channel varies from 0.04 to 0.05 ft./ml., computed velocity is quite high (2.7 ft./sec. to 4.1 ft./sec. at bankfull). The channel is small and cross-sectional area varies little (2.1 to 2.8 ft.^{2}), and corresponding Q_{bc} was small (5.5 ft.^{3}/sec. to 10.1 ft.^{3}/sec., with a mean value of 8.2 ft.^{3}/sec.). No pool and riffle sequence was apparent, and the thalweg was quite well-defined in the ill-defined banks. Hydraulic variables appeared to change little throughout the study segment.

Channel Segment 4 (Figs. 31 and 32)

Channel segment 4 is similarly non-sinuous (1.1) and shallow (R varies from 0.3 to 0.5, and the form ratio from 10.9 to 25.0). Values of n are about the same as for segment 3, roughness here being a result of the gravel bars evident along the channel floor rather than of scattered vegetation, as on the banks of channel 3. Slope S is quite high and varies from 0.03 to 0.05 ft./ml. throughout the study section, although there is generally little vegetation along the channel. Computed bankfull velocity is in general slightly higher than channel 3 due to the higher gradient, and is highest for cross-sections (i) and (vi), the latter being the location of a gravel bar. As with channels 1 and 3, no systematic pool and riffle sequence can be noted. The channel is tending to braid in places, especially adjacent to cross-sections (ii) to (iv) and (ix). Typically enlarged cross-sectional areas result with a discernable increase in width over depth at these locations. Cross-

Fig. 31

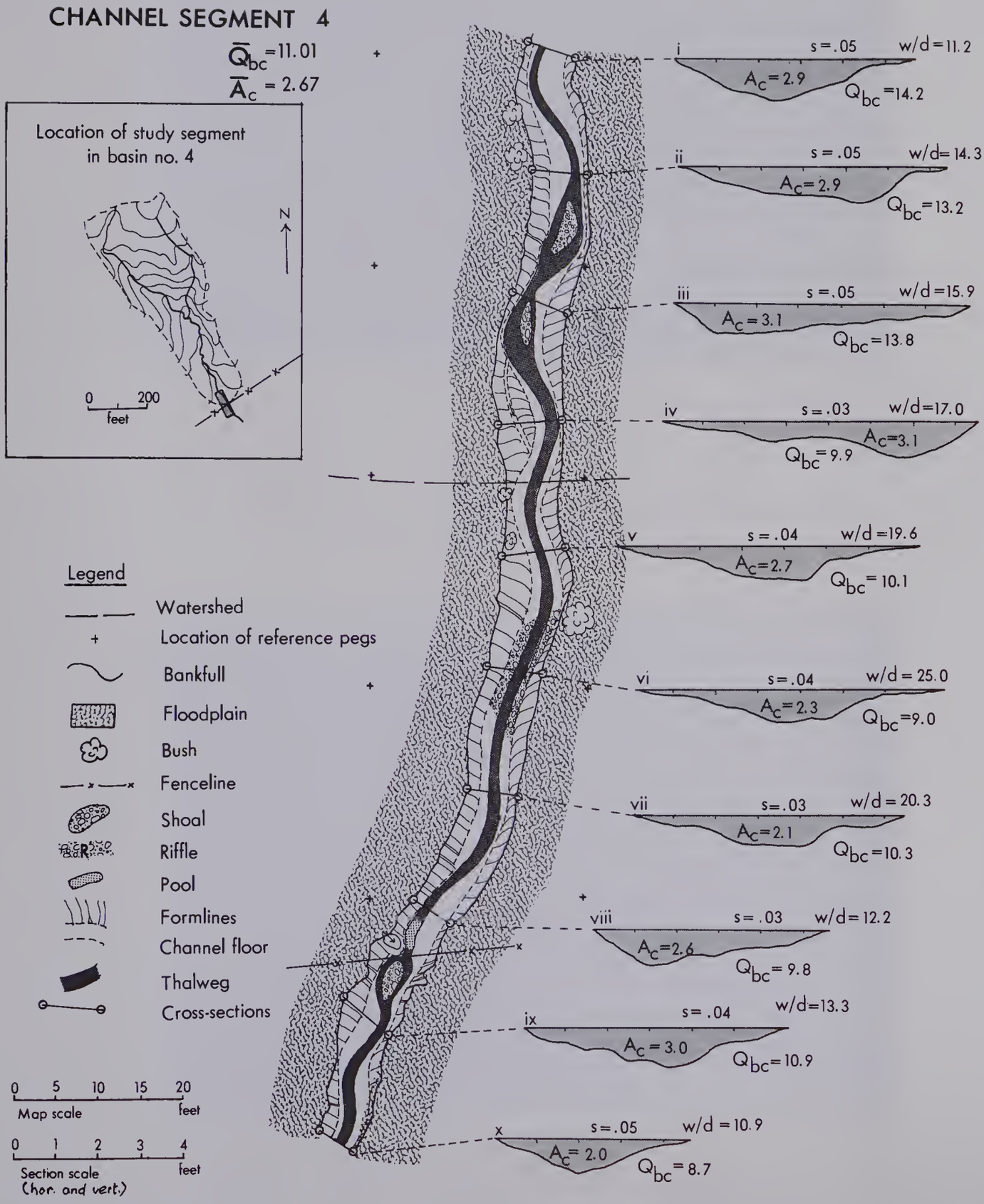

CHANNEL SEGMENT 4
$\bar{Q}_{bc}$=11.01
$\bar{A}_c$ = 2.67
Location of study segment in basin no. 4
N
0 200 feet
Legend
Watershed
Location of reference pegs
Bankfull
Floodplain
Bush
Fenceline
Shoal
Riffle
Pool
Formlines
Channel floor
Thalweg
Cross-sections
0 5 10 15 20 Map scale feet
0 1 2 3 4 Section scale (hor. and vert.) feet
i s=.05 w/d=11.2 A_c=2.9 Q_{bc}=14.2
ii s=.05 w/d=14.3 A_c=2.9 Q_{bc}=13.2
iii s=.05 w/d=15.9 A_c=3.1 Q_{bc}=13.8
iv s=.03 w/d=17.0 A_c=3.1 Q_{bc}=9.9
v s=.04 w/d=19.6 A_c=2.7 Q_{bc}=10.1
vi s=.04 w/d=25.0 A_c=2.3 Q_{bc}=9.0
vii s=.03 w/d=20.3 A_c=2.1 Q_{bc}=10.3
viii s=.03 w/d=12.2 A_c=2.6 Q_{bc}=9.8
ix s=.04 w/d=13.3 A_c=3.0 Q_{bc}=10.9
x s=.05 w/d=10.9 A_c=2.0 Q_{bc}=8.7



Fig. 32a Basin 4, facing upstream from the fenceline. The effect of the indurated sandstone bands is apparent.



Fig. 32b Channel 4 at watershed intersection, facing upstream.

sectional area varies from 2.0 ft.^{2} to 3.1 ft.^{2}, and computed discharge from 9.8 ft.^{3}/sec. to 14.3 ft.^{3}/sec. Once again, little variation in hydraulic variables occurs throughout the section.

Channel Segment 5 (Figs. 32 and 33)

As with channels 1 to 4, channel segment 5 is cut into relatively cohesive silty deposits, but the bankfull line is better defined than either 3 or 4, with channel sides exhibiting gullies similar to those in channels 1 and 2. The form of the channel is shallower than in 3 or 4 (R varies from 0.3 to 0.6, and form ratio from 6.4 to 21.1), and the channel is also more sinuous (1.2) and varied.

Values of S are moderate to high (0.02 to 0.05 ft./ml.) being lowest at (v) (vi) and (x) of which (v) and (x) correspond to gravel bar locations. Here, and at the riffle bar section (iii) (S = 0.04) the form ratios are highest, corresponding to the usual width increase over the riffle. Cross-sectional areas are highest at the location of the pools (vi) and (vii) where the form ratios decrease to 12.5 and 10.6 respectively. The channel continues its narrow, deep form below the pools,possibly in response to higher slopes and cohesive vegetation above the channel banks here. The roughness coefficient, n is consistent along the channel, but was assigned a higher value at (vi) and (vii) where boulders on the channel sides artificially restrict the channel. The second pool location corresponds to the location of these boulders.

Gullies on the channel banks cause the "stepped" appearance of the channel sections especially (vii) to (ix), as in channels 1 and 2. It must be concluded that rills developing in response to short

Fig. 33

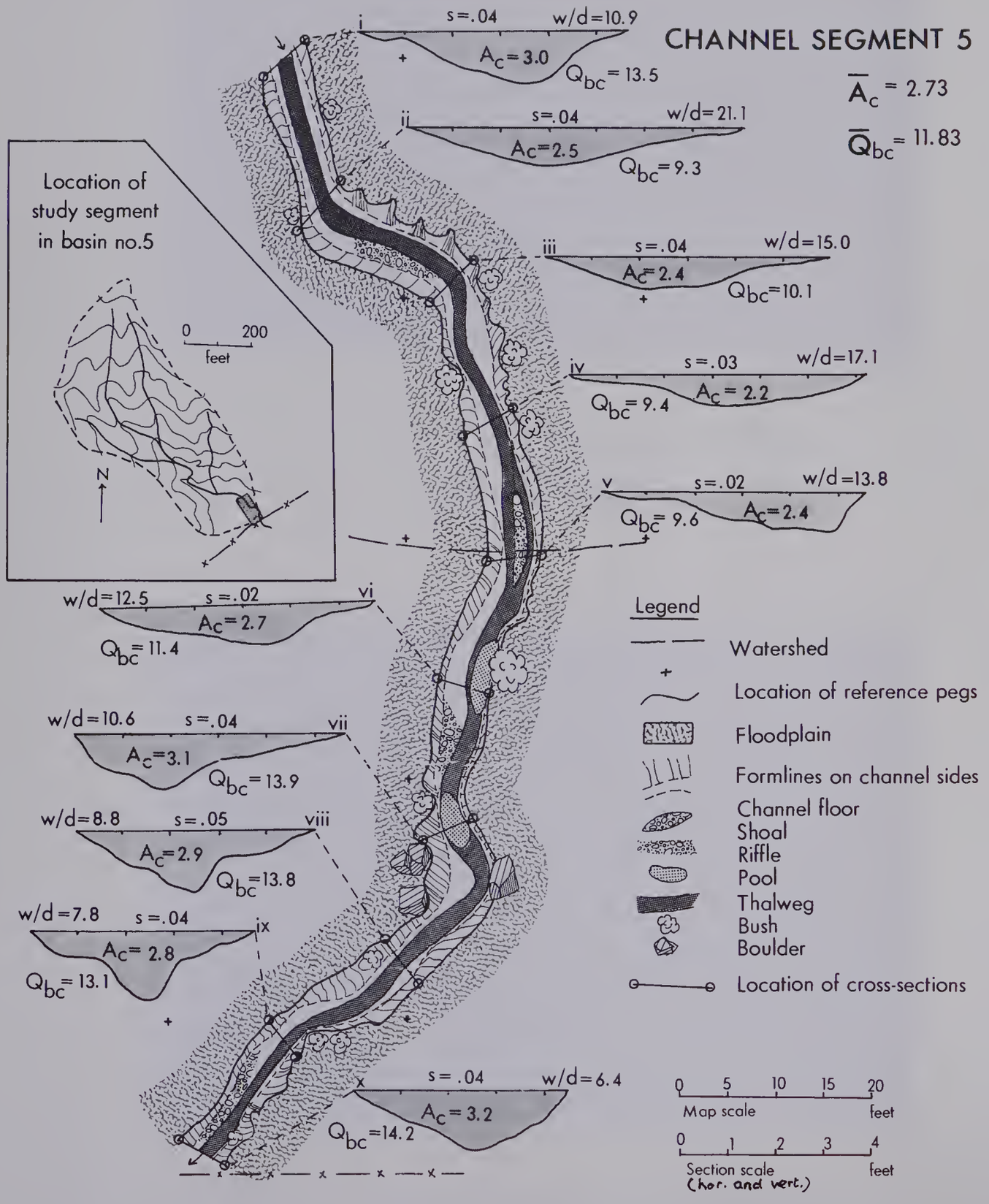

CHANNEL SEGMENT 5
$\overline{A}_c$ = 2.73
$\overline{Q}_{bc}$ = 11.83
Location of study segment in basin no.5
0 200 feet
N
i s=.04 w/d=10.9 A_c=3.0 Q_{bc}=13.5
ii s=.04 w/d=21.1 A_c=2.5 Q_{bc}=9.3
iii s=.04 w/d=15.0 A_c=2.4 Q_{bc}=10.1
iv s=.03 w/d=17.1 A_c=2.2 Q_{bc}=9.4
v s=.02 w/d=13.8 A_c=2.4 Q_{bc}=9.6
vi w/d=12.5 s=.02 A_c=2.7 Q_{bc}=11.4
vii w/d=10.6 s=.04 A_c=3.1 Q_{bc}=13.9
viii w/d=8.8 s=.05 A_c=2.9 Q_{bc}=13.8
ix w/d=7.8 s=.04 A_c=2.8 Q_{bc}=13.1
x s=.04 w/d=6.4 A_c=3.2 Q_{bc}=14.2
Legend
Watershed
Location of reference pegs
Floodplain
Formlines on channel sides
Channel floor
Shoal
Riffle
Pool
Thalweg
Bush
Boulder
Location of cross-sections
0 5 10 15 20 Map scale feet
0 1 2 3 4 Section scale (hor. and vert.) feet



Fig. 34a Basin 5, showing channel emergence from the watershed.



Fig. 34b Channel 5 at D (25 feet below watershed intersection.) The effects of vegetation on channel form are well illustrated.

duration, high intensity storms or to falling stage conditions (when these form) are an additional channel-forming mechanism, and that A_c may not reflect Q_{bc} in these cases.

Computed bankfull discharge varies in response to the changing cross-sectional area since velocity computation reveals few differences along the channel (except (vii), below the boulder restriction where velocity increases locally). No consistent pool and riffle sequence, and in this channel, channel adjustment to varying hydraulic conditions is generally less clear than in other channels.

Channel Segment 6 (Figs. 35 and 36)

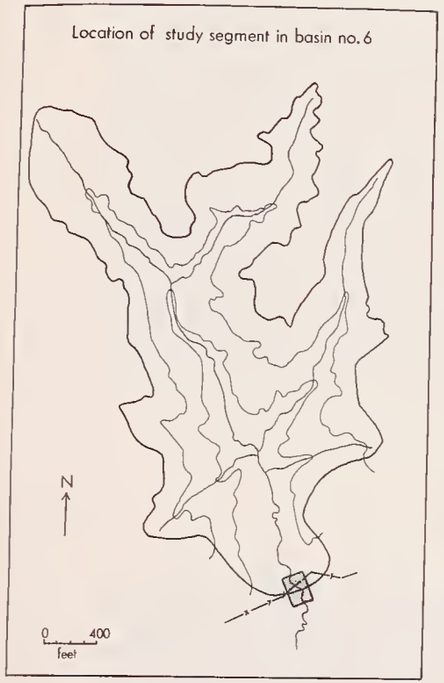
Channel segment 6 is by far the largest and most sinuous (1.4) of the ten channels studied, and in addition flows within the remnants of what appears to be a terrace. The feature, however, is difficult to explain. Digging into the margin of the terrace revealed no obvious junction between the present flood plain and the upper feature, and there is a possibility that this form may be the termination of alluvial fans from peripheral tributaries. On mapping, the form resembled a terrace, but examination of the other basins failed to reveal any similar feature; although the continuation of this form is apparent in the upstream sub-basin segment 7. The present bankfull level is at a consistent depth below the terrace, and the channel is tortuous within its confines. Both upper terrace and present flood plain are vegetated.

The present channel offers little obstruction to flow along straight reaches, and slopes are gentle, S varying from 0.001 to 0.02, and calculated velocity at bankfull varies in response to slope, from 1.8 ft./sec. to 3.8 ft./sec. Pools and riffles are closely spaced (3.3w

Fig. 35 CHANNEL SEGMENT 6

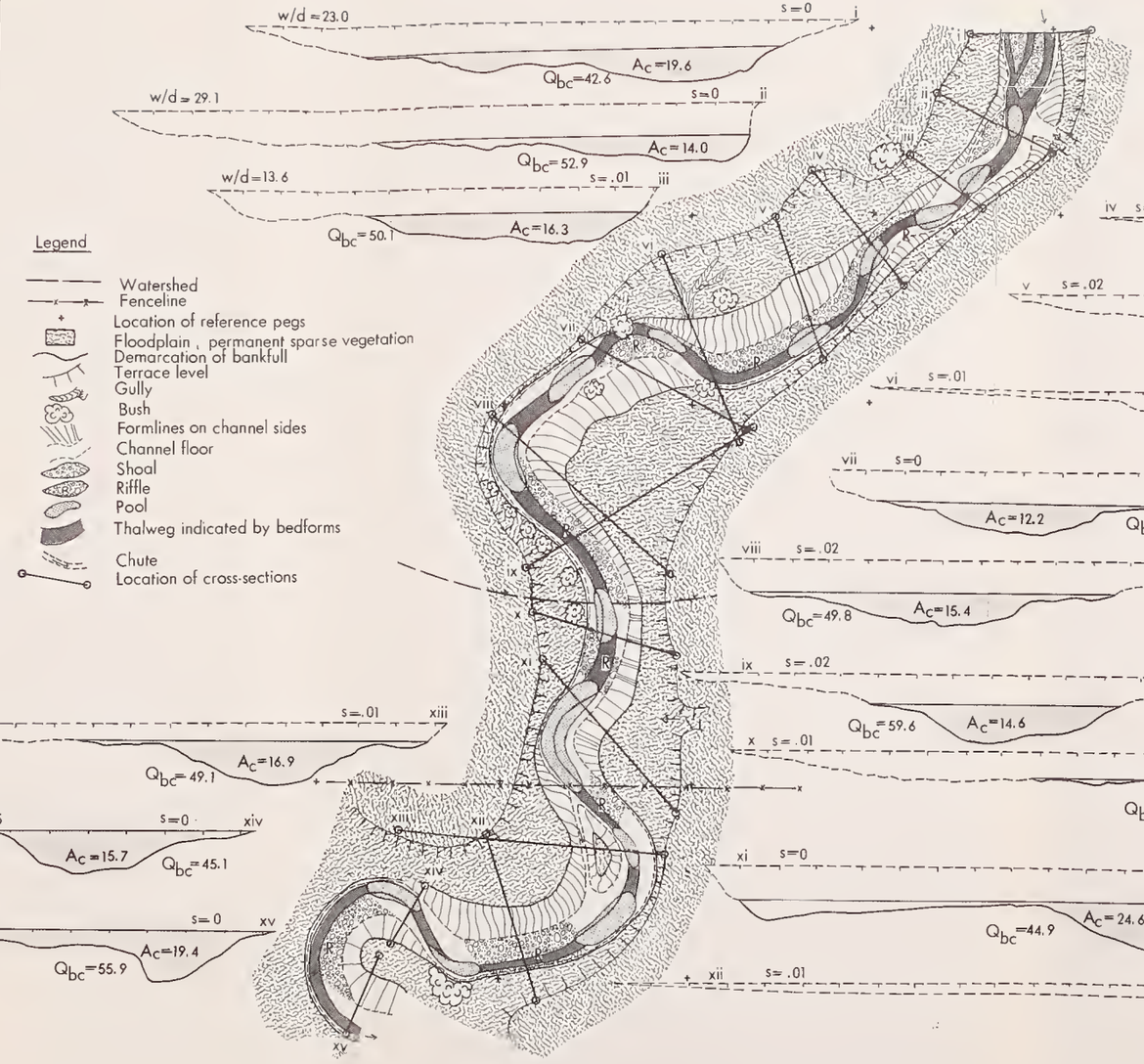
$\bar{A}_c = 17.35$

$\bar{Q}_{bc} = 51.01$



Legend

- Watershed
- Fenceline
- Location of reference pegs
- Floodplain, permanent sparse vegetation
- Demarcation of bankfull
- Terrace level
- Gully
- Bush
- Formlines on channel sides
- Channel floor
- Shoal
- Riffle
- Pool
- Thalweg indicated by bedforms
- Chute
- Location of cross-sections



i s=0 w/d=23.0 A_c=19.6 Q_{bc}=42.6

ii s=0 w/d=29.1 A_c=14.0 Q_{bc}=52.9

iii s=.01 w/d=13.6 A_c=16.3 Q_{bc}=50.1

iv s=.02 w/d=10.5 A_c=20.7 Q_{bc}=66.1

v s=.02 w/d=26.5 A_c=16.1 Q_{bc}=45.3

vi s=.01 w/d=30.7 A_c=13.8 Q_{bc}=50.5

vii s=0 w/d=21.6 A_c=12.2 Q_{bc}=46.1

viii s=.02 w/d=25.0 A_c=15.4 Q_{bc}=49.8

ix s=.02 w/d=12.2 A_c=14.6 Q_{bc}=59.6

x s=.01 w/d=10.2 A_c=14.9 Q_{bc}=49.7

xi s=0 w/d=82.4 A_c=24.6 Q_{bc}=44.9

xii s=.01 w/d=21.2 A_c=25.8 Q_{bc}=57.5

xiii s=.01 w/d=20.8 A_c=16.9 Q_{bc}=49.1

xiv s=0 w/d=18.5 A_c=15.7 Q_{bc}=45.1

xv s=0 w/d=21.0 A_c=19.4 Q_{bc}=55.9

0 10 20 30 40 feet
Map scale

0 2 4 6 8 feet
Section scale
(hor. and vert.)



Fig. 36a Location of Channel study segment at the point of watershed emergence, Basin 6.



Fig. 36b Channel 6 at (vi). Tape is level with the terrace, present flood plain is some 2 feet below this. Facing upstream.

on average) and cross-sectional area at pools is in general larger than that at riffles, although where braiding occurs (i) and where a chute was developed (xii) the cross-sectional area is artificially enlarged in response to the change in hydraulic control mentioned previously. The shortened repeated spacing of pools and riffles may be a result of the restriction of the upper terrace (if this is what the feature is). Dury (1967) has described this phenomenon in the Hawkesbury River, New South Wales, where he distinguishes "short-series pools" from "long-series pools" as the basis of control from a previous channel remnant. The cross-sectional area of the channel varies from 14.0 ft.^{2} at (ii) to 25.8 ft.^{2} at the chute and island (xii). Q_{bc} varies between 42.6 and 66.1 ft.^{3}/sec. with a mean value of 51.0 ft.^{3}/sec.

Channel Segment 7 (Figs. 37 and 38)

Channel segment 7 was located just below the confluence of two fourth order basins with the main channel, and at this location the channel is almost straight (sinuosity = 1.0) and moderately steep (0.01 to 0.04 ft./ml. at the sections). As with channel 6, the present channel flows within an apparent terrace level which in this case parallels the bankfull margins.

Particularly noticeable on this channel section is the progressive increase in cross-sectional area along the study length, from 10.72 ft.^{2} at (i) to 17.92 ft.^{2} at (x). This occurs only in this channel, and is due to the fact that basin 7 is a sub-basin of 6, and whereas all other channels emerge onto an alluvial fan after leaving the study section, the stream in channel segment 7 continues to enlarge downstream after cross-section (x). Despite this trend, the form ratio of the

Fig. 37

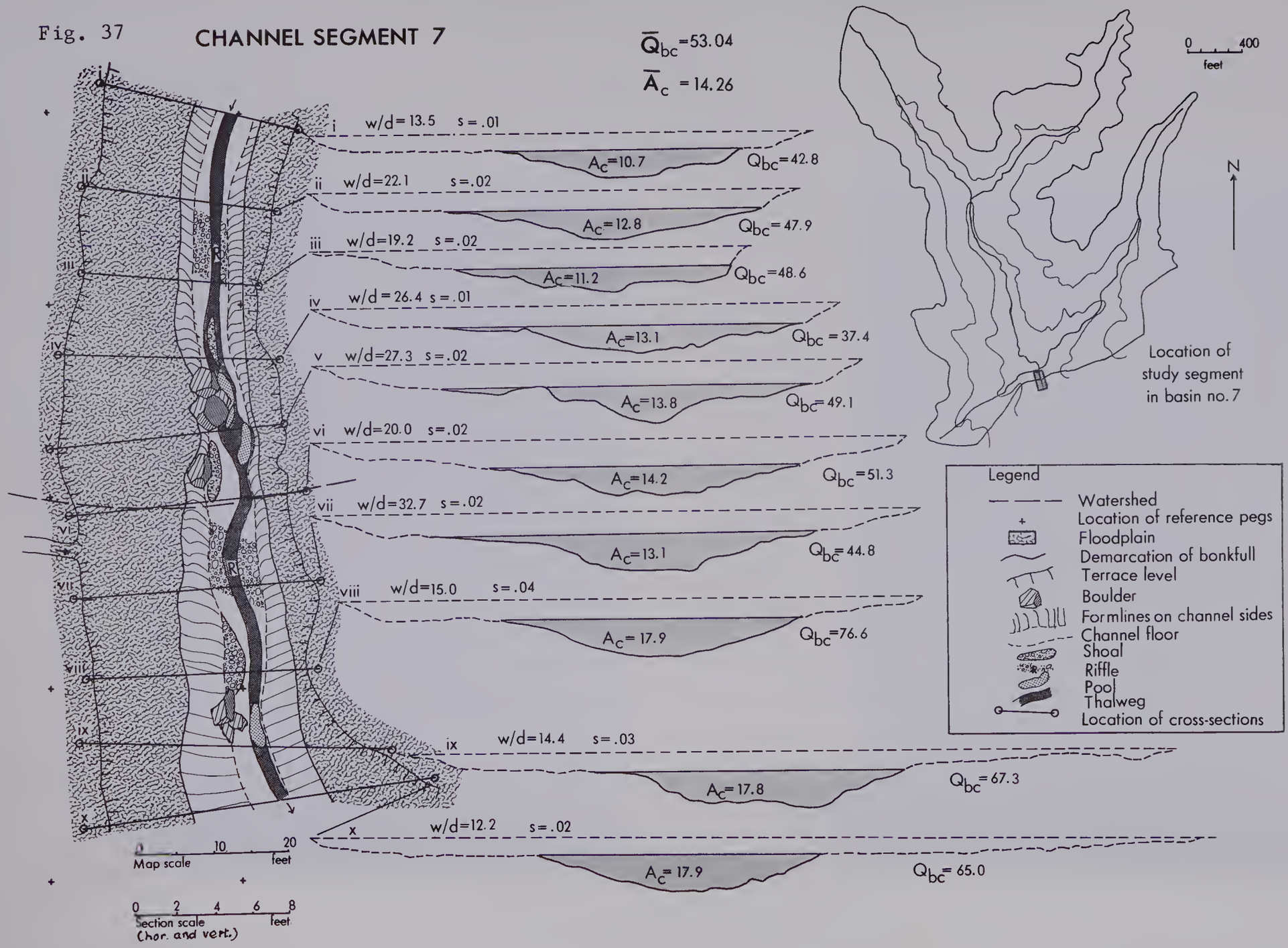

CHANNEL SEGMENT 7
$\overline{Q}_{bc}$=53.04
$\overline{A}_c$ = 14.26
i w/d=13.5 s=.01
A_c=10.7 Q_{bc}=42.8
ii w/d=22.1 s=.02
A_c=12.8 Q_{bc}=47.9
iii w/d=19.2 s=.02
A_c=11.2 Q_{bc}=48.6
iv w/d=26.4 s=.01
A_c=13.1 Q_{bc}=37.4
v w/d=27.3 s=.02
A_c=13.8 Q_{bc}=49.1
vi w/d=20.0 s=.02
A_c=14.2 Q_{bc}=51.3
vii w/d=32.7 s=.02
A_c=13.1 Q_{bc}=44.8
viii w/d=15.0 s=.04
A_c=17.9 Q_{bc}=76.6
ix w/d=14.4 s=.03
A_c=17.8 Q_{bc}=67.3
x w/d=12.2 s=.02
A_c=17.9 Q_{bc}=65.0
0 400 feet
N
Location of study segment in basin no. 7
Legend
Watershed
Location of reference pegs
Floodplain
Demarcation of bankfull
Terrace level
Boulder
Formlines on channel sides
Channel floor
Shoal
Riffle
Pool
Thalweg
Location of cross-sections
0 10 20 feet
Map scale
0 2 4 6 8 feet
Section scale (hor. and vert.)



Fig. 38a Channel segment 7 in basin 6.

Facing upstream.



Fig. 38b Intersection of watershed

and channel 7.

channel shows no correspondence from (i) to (x) being highest at (vii) and lowest at (x). This lack of trend is a function of the varying width of the channel, since an examination of depth (d) downstream shows a progressive increase. Between sections (v) and (vi) a pile of boulders causes local eddying resulting in three small pools, where the channel takes on a very circular form. However, since no cross-section was measured here, the hydraulic radius values most indicative of circularity are those for cross-sections (viii) (ix) and (x), which imply that efficiency is increasing in the lower sections. The channel below (vi), however, is strewn with quite large boulders, causing n values to be higher. Thus the computed velocity shows no increase throughout the section, and as a result Q_{bc} values increase in correspondence to the increasing cross-sectional area. This varies from 37.4 ft.^{3}/sec. at (iv) to 76.6 ft.^{3}/sec. at (viii) with a mean value of 53.1 ft.^{3}/sec.

Channel Segment 8 (Figs. 39 and 40)

Number 8, the smallest channel emerges from a steep narrow watershed, and consequently the first three cross-sections are located on quite steep slopes (0.06 to 0.04 ft./ml.) whereas the last few have a gentler slope through the cross-section at bankfull (0.02 to 0.03 ft./ml.). The channel is quite sinuous (1.2) and the bank materials are cohesive similar to the other channels, but a bedrock outcrop occurs on the eastern bank in the upper section and causes a break of slope just above section (iv). Two small tributaries enter in the study segment but were considered to have a very minor effect on the cross-sectional characteristics of the channel. This was substantiated by an examination of the A_c values for (i) to (x) which showed no increase after the two

Fig. 39

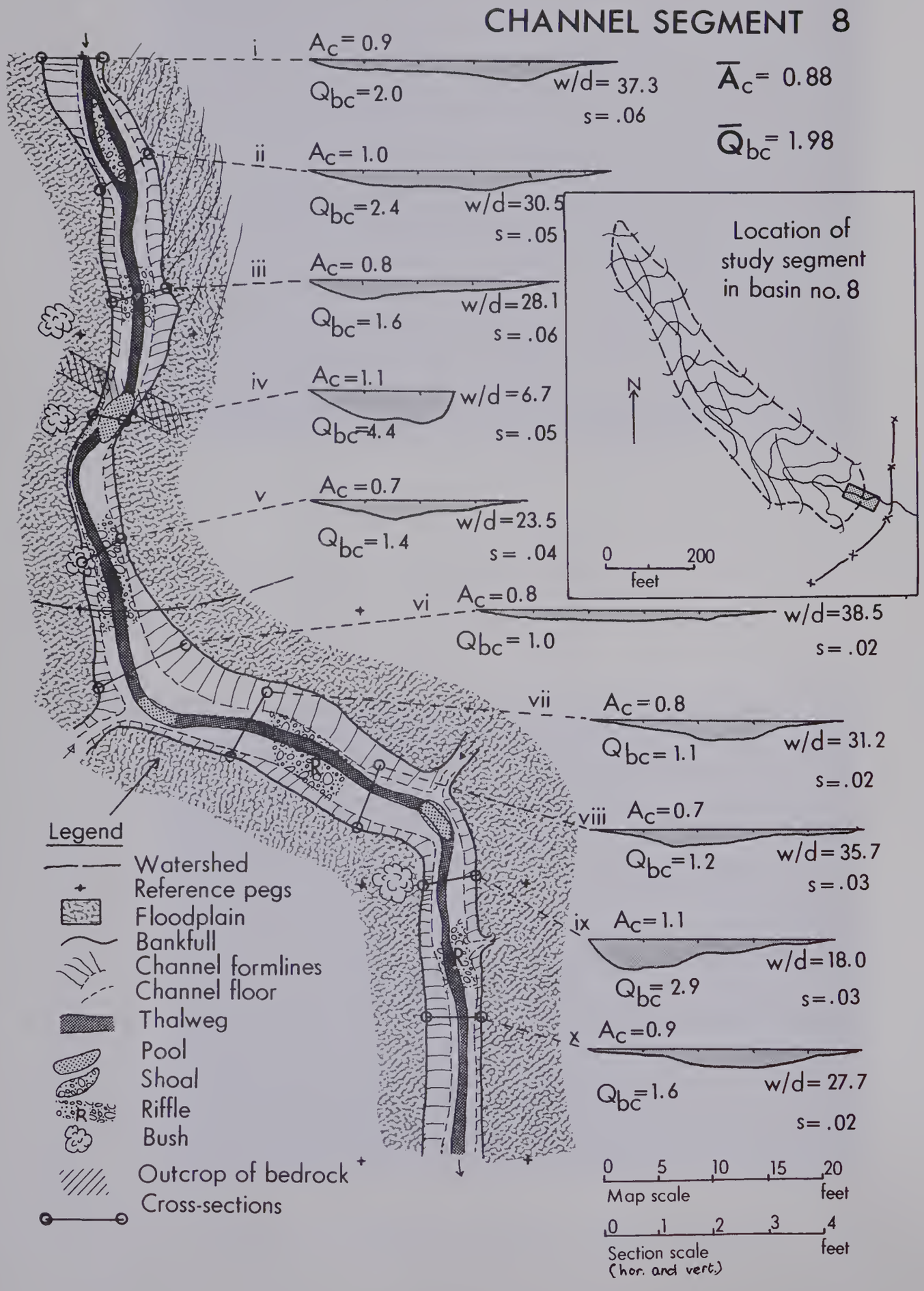

CHANNEL SEGMENT 8
$\bar{A}_c$ = 0.88
$\bar{Q}_{bc}$ = 1.98
i A_c = 0.9 Q_{bc} = 2.0 w/d = 37.3 s = .06
ii A_c = 1.0 Q_{bc} = 2.4 w/d = 30.5 s = .05
iii A_c = 0.8 Q_{bc} = 1.6 w/d = 28.1 s = .06
iv A_c = 1.1 Q_{bc} = 4.4 w/d = 6.7 s = .05
v A_c = 0.7 Q_{bc} = 1.4 w/d = 23.5 s = .04
vi A_c = 0.8 Q_{bc} = 1.0 w/d = 38.5 s = .02
vii A_c = 0.8 Q_{bc} = 1.1 w/d = 31.2 s = .02
viii A_c = 0.7 Q_{bc} = 1.2 w/d = 35.7 s = .03
ix A_c = 1.1 Q_{bc} = 2.9 w/d = 18.0 s = .03
x A_c = 0.9 Q_{bc} = 1.6 w/d = 27.7 s = .02
Location of study segment in basin no. 8
N
0 200 feet
Legend
Watershed
Reference pegs
Floodplain
Bankfull
Channel formlines
Channel floor
Thalweg
Pool
Shoal
Riffle
Bush
Outcrop of bedrock
Cross-sections
0 5 10 15 20 Map scale feet
0 1 2 3 4 Section scale (hor. and vert.) feet



Fig. 40a Channel 8 at its emergence

from basin 8.



Fig. 40b Channel 8 at (viii), facing upstream.

tributaries joined the main channel. The channel was well defined above (vi), but locating the bankfull line proved problematical on (vii) to (x) which were extremely wide and shallow. Cross-sections all exhibit high form ratios, varying from 18.0 to 38.5, except for (iv) - 6.7 - which was constricted laterally by the rock outcrop and had consequently developed vertically. Cross-section areas varied little, around 1 ft.^{2}, and velocity decreased downslope in response to gentler gradient. Q_{bc} varied from 1 ft.^{3}/sec. (where the channel was wide and flat and thus experienced drag on flow) to 2.9 ft.^{3}/sec. at (ix). The mean value was 1.98 ft.^{3}/sec. Pools and riffles were observed at 4.7w recurrent interval.

Channel Segment 9 (Figs. 41 and 42)

Channel segment 9 is not sinuous (1.0) and has relatively few obstructions to flow. Thus n values were accordingly low along the channel, which is quite wide and shallow but presents quite a smooth form in cross-section. The effect of gullying on the channel sides at flows less than bankfull is particularly obvious at (viii) where a well-developed gully on the channel walls has eaten back into the flood plain artificially increasing 2, and resulting in a very high form ratio (59.1). Form ratios vary downstream from 10.5 to 59.1, and slopes through the cross-sections are highest in the upper channel and lowest on the wider channel downstream. Velocities are generally high due to the smooth profile of the channel, and are maximum where the slopes through the sections are highest. Q_{bc} varied from 19.9 ft.^{3}/sec. at (viii) to 39.3 ft.^{3}/sec. at bankfull. Mean value was 27.1 ft.^{3}/sec.

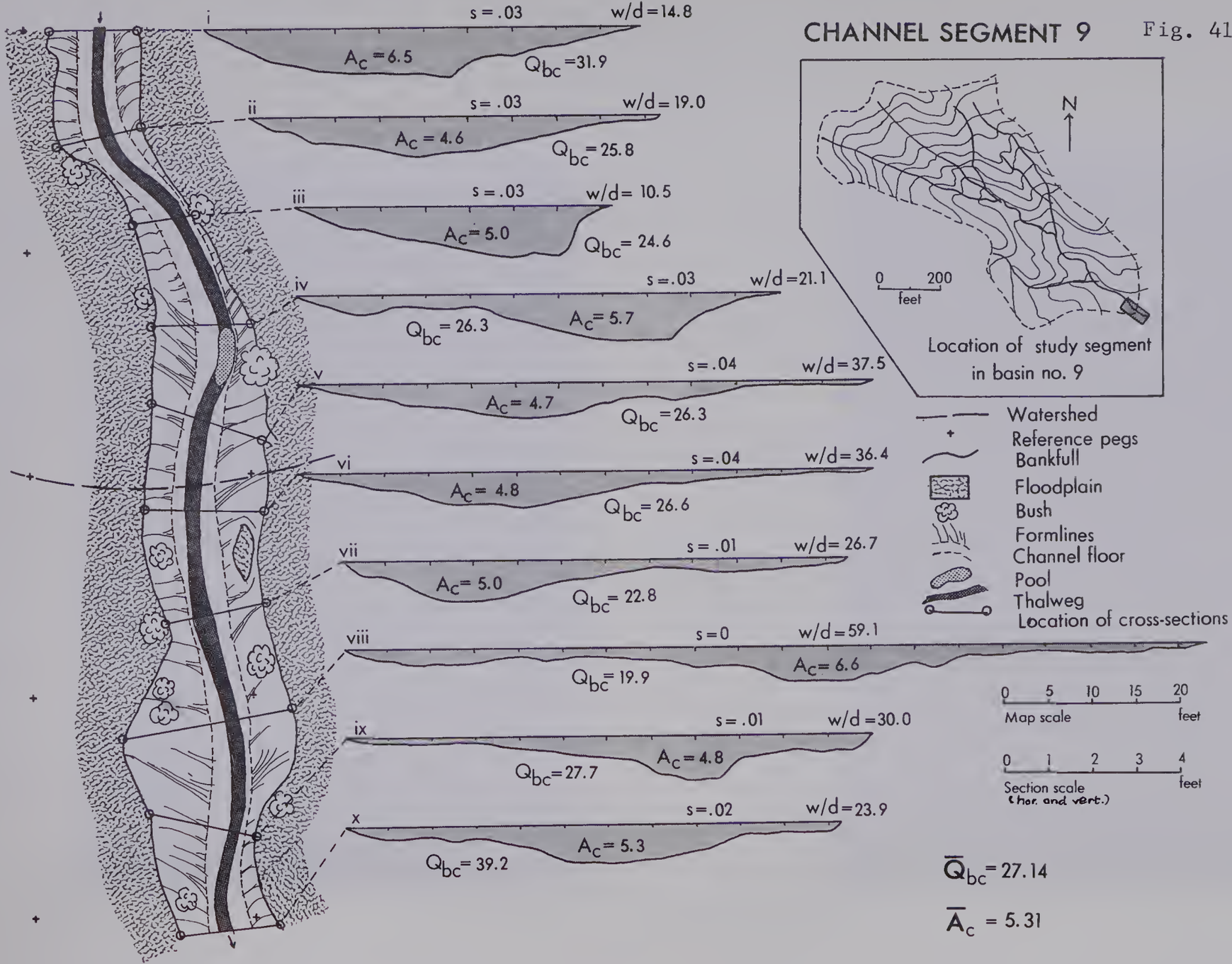

CHANNEL SEGMENT 9
Fig. 41
i
s = .03
w/d = 14.8
Ac = 6.5
Qbc = 31.9
ii
s = .03
w/d = 19.0
Ac = 4.6
Qbc = 25.8
iii
s = .03
w/d = 10.5
Ac = 5.0
Qbc = 24.6
iv
s = .03
w/d = 21.1
Qbc = 26.3
Ac = 5.7
v
s = .04
w/d = 37.5
Ac = 4.7
Qbc = 26.3
vi
s = .04
w/d = 36.4
Ac = 4.8
Qbc = 26.6
vii
s = .01
w/d = 26.7
Ac = 5.0
Qbc = 22.8
viii
s = 0
w/d = 59.1
Qbc = 19.9
Ac = 6.6
ix
s = .01
w/d = 30.0
Qbc = 27.7
Ac = 4.8
x
s = .02
w/d = 23.9
Ac = 5.3
Qbc = 39.2
N
0 200
feet
Location of study segment in basin no. 9
Watershed
Reference pegs
Bankfull
Floodplain
Bush
Formlines
Channel floor
Pool
Thalweg
Location of cross-sections
0 5 10 15 20
Map scale
feet
0 1 2 3 4
Section scale
(hor. and vert.)
feet
$\overline{Q}_{bc} = 27.14$
$\overline{A}_c = 5.31$



Fig. 42a Basin 9, showing location of channel at watershed emergence.



Fig. 42b Channel 9 at (viii), facing upstream. Field assistant shown relocating reference peg C.

Channel Segment 10 (Figs. 43 and 44)

Although the thalweg of channel segment 10 is straight, (sinuosity = 1.0) the pattern of the channel is braided between (i) and (vii) and cross-sections are correspondingly wide and shallow in this stretch. Form ratios vary from 12.5 below the braid to 33.3 at the braided section, and R drops from 0.57 below to 0.35 at the braid. The banks of the channel are in cohesive material similar to other channels but are gullied in parts between (iii) and (iv) and (vi) and (vii) in particular. In the lower section the channel narrows and deepens, and is considerably roughened along its bed by a gravel riffle over which there is an increase in slope. Associated with the braided reach, slope through cross-section (v) is high compared to that above and below, in agreement with observations made by Rubey (1952) and Leopold and Wolman (1957) about the change in hydraulic conditions associated with braiding. Through the braided reach and below, channel roughness and cross-sectional area correspondingly increase. Velocities vary in response to changing slope.

The channel tends to enlarge downstream with a few variations, and discharge increases from 9.3 ft.^{3}/sec. to 17.8 ft.^{3}/sec. with a mean value of 14.4 ft.^{3}/sec.

D. Conclusions

Although it is difficult to isolate direct cause-and-effect relationships between formative processes and the channel form and pattern, several observations can be made from the examination of the ten channel segments studied.

(1) All channels are cut into relatively fine-grained alluvial deposits.

Fig. 43 **CHANNEL SEGMENT 10**

$\overline{Q}_{bc}$ = 14.38

$\overline{A}_c$ = 3.76

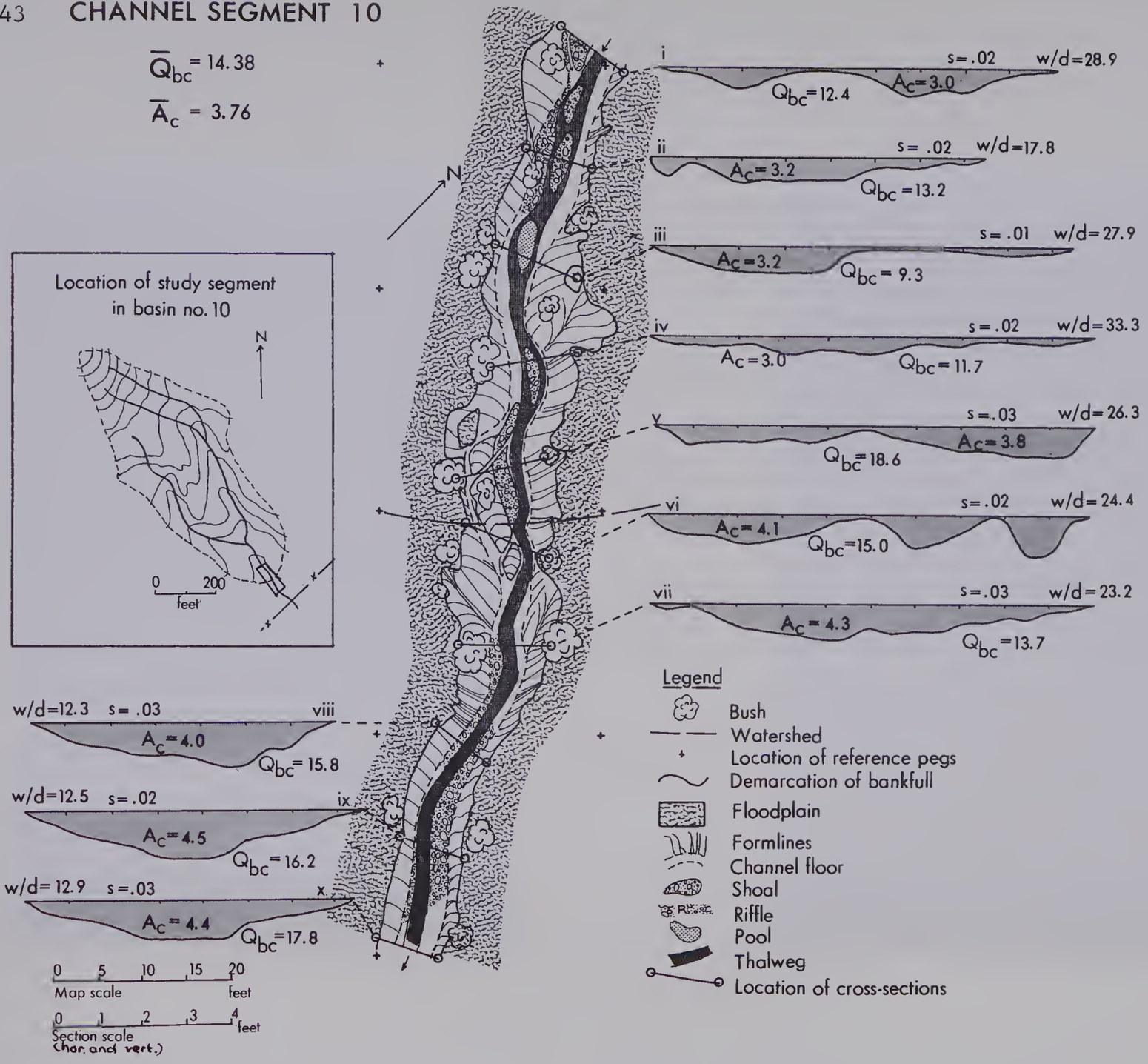




Fig. 44a Basin 10, showing channel at its watershed emergence.



Fig. 44b Channel 10 at (ix), facing upstream.

(2) The flood plain is sparsely vegetated and only some larger sage bushes cause changes in the channel pattern. Channel width adjustments are thus easily made.

(3) Gullying on channel sides is caused by direct rainfall or by run-off following bankfull; and wash at discharges less than bankfull. These gullies form independently of bankfull discharge values and can be accredited to the climate, the ephemeral nature of the channels, and the relatively easily eroded bank material.

(4) Form ratio and hydraulic radius are largely dependent on the nature of the bed and banks, but several other factors can cause variations.

(a) A sharp increase in slope above the channel segment (e.g. number 1).

(b) A change in channel pattern. Braiding and riffles cause an increase in width over depth; pools and channel restrictions cause an increase in depth over width.

(c) Chutes cause aggradation of the main channel due to their ability to divert flow at flood stage. Thus, S decreases, A_c and w increase.

(d) There is a width increase associated with bends.

(5) Channels are similarly rough except where highly sinuous, or have vegetation on channel banks or gravel and boulders within the bed.

(6) Computed velocities (using the Manning equation) are within similar ranges and appear credible.

(7) The average cross-sectional area of each of the channels A_c, and the average bankfull discharge, Q_{bc}, for each channel may be related to the area of the basin (A_s), the length of the main channel in the basin (L_s) and its gradient (GR). These relationships will be examined in the next chapter.

CHAPTER IV

CORRELATIONS AND CONCLUSIONS

In order to establish the validity of the designed model, it is necessary to relate the morphometric variables to those existing within the channel.

A. Relationships between channel and basin.

During the last thirty years a large proportion of geomorphic literature has been devoted to quantification of the relationships existing within the drainage net. Since 1945, when Horton outlined three basic laws of drainage net composition, number, length and area drained by streams of increasing order have been studied in many basins and found to approximate a geometric series, and the ratios in each case have been given geomorphic significance.

Other laws of basin morphometry have been outlined, notably by Schumm (1956b), who worked within a small basin in the Perth Amboy badlands, New Jersey. Geometric properties including relief ratio, elongation ratio, gradient of the main channel, and drainage density were successfully related to the value of the hypsometric integral (Schumm, 1956b, p. 617), thus providing a link between the two-dimensional erosional development of the drainage system.

One morphometric relationship which represents a synthesis of the Hortonian analysis is that between the size of a drainage basin (A_d), and the length (L_d) and gradient (GR) of its main channel. Hortonian analysis provides evidence that the length of streams of increasing order O increase geometrically with order (Horton, 1945), that area

drained by streams of increasing order (A_o) increases geometrically with O (Schumm, 1956b, after Horton, 1945) and also that gradient (GR) of streams of increasing order O increases geometrically with order (Horton, 1945). Thus Hack (1957) suggested that for any basin the relationships L_d with A_d, L_d with GR, and A_d with GR must be expected in a downstream direction, and probably between basins closely similar both geologically and climatically. For 90 study points on the Shenandoah valley, and adjacent mountains in Virginia, Hack related both A_d and L_d to the Hortonian analysis, and thence fitted values to a line whose equation is:

$$L_d = 1.4\ (A_d)^{0.6} \qquad (1)$$

where L_d and A_d are in miles, and miles^2 respectively. Very little scatter occurred from the fitted line. Hack also related GR and size of bed material to L_d. He concluded (p. 63) that "...length...and drainage area are interdependent quantities such that one changes with respect to the other at a rate that appears to be roughly uniform over large areas." Brush (1961) has applied equation (1) to selected streams in Pennsylvania with some success.

In a later study, Morisawa (1962) examined this relationship in the Allegheny plateau, and pointed out that for basins not elongating downstream the value of the exponent should be theoretically 0.5, but that downstream basin elongation made 0.6 a better approximation in most cases. Gray (1961) fitted values of basin length and area to the line:

$$L_d = 1.4\ (A_d)^{0.57} \qquad (2)$$

where L_d and A_d are in miles, and miles^2 respectively and also examined the relationship between GR and other basin variables. Leopold, et al (1964) studied equation (1) on various basins, including some major rivers of the world, and stated that the fact that,

> "...the exponent is not 0.5 but 0.6 to 0.7 indicates that with increasing basin size, drainage basins elongate,; length increases faster than width," and that "...this is true through a wide spectrum of basin size." (p. 145).

Smart and Surkan (1962) examined the value of this relationship, pointing out that "...the main stream length...has been previously used as a single parameter to replace the s quantities $\bar{L}_1$, $\bar{L}_2$, $\bar{L}_3$,....L_s in any characteristic stream system." (p. 964). They noted, however, that main stream sinuosity appears to have a small but significant variation with area, and that this can make a "significant contribution" to the deviation from 0.5 in the exponent of the L_d with A_d equation along with basin elongation. Therefore, scientifically L_d does not seem to be the ideal parameter to use for characterisation of a drainage basin.

Besides the relationship L_d with A_d, these two variables have also been related to bankfull discharge, Q_b. In particular, Q_b has been related to A_d by the equation:

$$Q_b \propto A_d^{0.75} \qquad (3)$$

where Q_b and A_d are in ft.^{3}/sec. and miles^2 respectively. Leopold, et al (1964, p. 251) have said that this;

> "...varies somewhat amongst different climates, but can be closely specified for many basins. The exponent of drainage area varies from 0.65 to 0.80, a general average being 0.70 to 0.75," and that "...the value of these exponents need not be considered as wholly unexplained empiricism."

Since A_d and L_d increase with O, and Q_b largely controls channel form, it can be expected that hydraulic variables can be related through Q_b to A_d, L_d and hence provide a link to the basin ordering. In studying ephemeral channels in New Mexico, Leopold and Miller (1956, p. 1) noted that;

> "...stream order is related to stream length, number of streams, drainage area, and discharge by simple exponential functions. The relation of discharge to width, depth, velocity and slope and other hydraulic factors can be approximated by simple

> power functions. Thus any pair of these factors is related by exponential or power functions."

In their study, and also those of Leopold (1953), Leopold and Maddock (1953) and Brush (1961), the nature of channel geometry changes with variations in discharge was outlined. Since $Q = w.\ d.\ V$, then obviously $w = aQ^b$, $d = cQ^f$ and $V = kQ^m$, where b, f, and m are exponents whose product is unity, and a, c, and k are constants whose product is unity. Thus w, d, and V can be regressed with Q at-a-station for variable Q, and in a downstream direction for one defined value of Q, usually Q_b. The exponents f and m vary considerably for different climatic regimes, but b has been found to approximate closely 0.5 in most of the cases studied; i.e. $w \alpha Q_b^{0.5}$ (4). Suspended sediment load W_1 can be similarly found to correlate well with Q_b, downstream and at-a-station. ($W_1 = pQ_b^{1.3}$, Leopold and Miller, 1956, p. 15).

These studies are comparable to Brush's work (1961) in Pennsylvania in which a comprehensive study of channel slope (GR) mainstream length (L_d) and basin area (A_d) successfully established connections within these basin variables, and between these and channel geometry variables in a downstream direction. Using the measured channel width and depth at various guaging stations, velocity at bankfull, V_b, was calculated using the Manning equation and correlated with discharge at the 2.3-year recurrence interval at each station. Power functions were derived for w, d, and V (b, f, and m) with $Q_{2.3}$ in the same manner as did Leopold and Maddock (1953). Bed material size was also correlated with GR and L_d. An interesting aspect of Brush's work is that 16 of the guaging stations were on different streams in similar geological environments, and he found that values obtained at these points could be considered as different

study sections on one hypothetical stream.

Since a basin and its channels form an interacting system, it is not surprising that many of the measurable forms of both channel and basin are exponentially related in this way. Current quantitative fluvial geomorphology is largely directed towards further documentation of these relationships, particularly in varying climatic and lithological environments so that the nature of landforming processes can be further elucidated.

(i) Relationships studied at Steveville

In Chapter I the mode of basin evolution in the study area was outlined by means of hypsometric curves and longitudinal stream profiles. Stream gradient GR, was found to lower with increasing basin size, A_s, at a consistent rate once the streams had breached the prairie level, whereas those developing on the scarp face showed no trend, being variously steep and controlled by lithologic variations on the retreating face.

In Chapter II the planimetric consistency of the drainage net was illustrated using Hortonian analysis, certain deviations being explicable in terms of the evolutionary model. In addition, the relationship between the average area drained by basins of a given order O ($\bar{A}_o$) for each of the stream orders was regressed with average length of streams of a given order ($\bar{L}_o$) and the exponent 0.4 was noted and shown to correspond as expected to the exponent in the regression L_s with A_s. This establishes the fact that the channel length with drainage area relationship can be carried through the drainage net in the same manner in each basin, right up to the first order basins. It is interesting to note that the exponent is lower than predicted in either (1) or (2). The theore-

tical value of 0.5 or above would be obtained if the point to the far right of the regression line (Appendix B) was not pulling the line down to a lower gradient. This is the sixth order length of basin 7 again, it may have been more meaningful to omit this point from regression analyses. Nevertheless, L_s with A_s does not have a higher exponent, and leads to the suggestion that perhaps a low exponent is characteristic of semiarid climates, i.e. area increases far more rapidly than in humid climates for a given length of main channel because peripheral gullies develop along the channel which may be almost as large as the main basin, but which are not accounted for in the L_s value, only in the A_s value. This seems quite feasible considering the non-competitive environment for dominant basins and their tendency to expand laterally rather than vertically once the prairie level has been breached.

In order to relate the morphology of the channels as they emerge from their watersheds to the morphometric analyses of the basin, the mean width, w, and mean cross-sectional area, A_c, were measured for each of ten channel segments, and the mean value of the bankfull discharge, Q_{bc}, calculated (Chapter III). These three channel characteristics can be related to the morphometric variables within their basins by means of the simple regressions proposed by Brush (1961), Hack (1957), Leopold and Miller (1956), Leopold and Maddock (1953), etc., and exponents in each regression compared to the values obtained by these workers wherever possible.

The study segments at Steveville are not on the same stream but on adjacent streams that have basins with comparable lithology and planimetric characteristics. Thus the study can be considered an extension of Brush's (1961) work in that it should be possible to regard

the study segments as if on one hypothetical stream, especially in the light of the consistency of the L_s and A_s relationship throughout the whole area.

In Fig. 45 the average cross-sectional area of the channel, A_c, can be seen to increase with the order of the basin, s. This relationship justified further regressions between basin and channel variables. Basin area, A_s, correlated very well with A_c, (Fig. 46) the best fit line to all points having an equation whose exponent is 1.4 ($A_s \propto A_c^{1.4}$), (5), which is significant at the 0.1% level. Main stream length, L_s, similarly correlates well with cross-sectional area (Fig. 47) the line being significant at the 1% level, and the exponent in this case being 0.5. The relationship between main channel gradient and cross-sectional area (Fig. 48) is not quite so clear, $GR \propto A_c^{0.4}$, (6), being significant at the 2% level only.

In each of these three regressions, the dotted line represents the trend of values for inhibited and dominant forms, and it is again evident that basins that are free to expand have a lower gradient than the norm, and a closer approximation to a straight line. The change occurs at stream lengths of about 1000 feet, or areas of approximately 250,000 $feet^2$, the largest size of basins of inhibited form. In other words, the channel enlarges more slowly with increasing basin size once the prairie level has been breached. This is difficult to visualise but long distances of overland flow allow for greater evaporation and seepage losses on these channels compared to the smaller ones. Thus a proportionately smaller channel is carved by the diminished flow.

A similar 'split' in the graphic trends for the larger basins in the graph of basin size with channel width (Fig. 49), although in

Fig. 45

Linear regression
Log. A_c with s

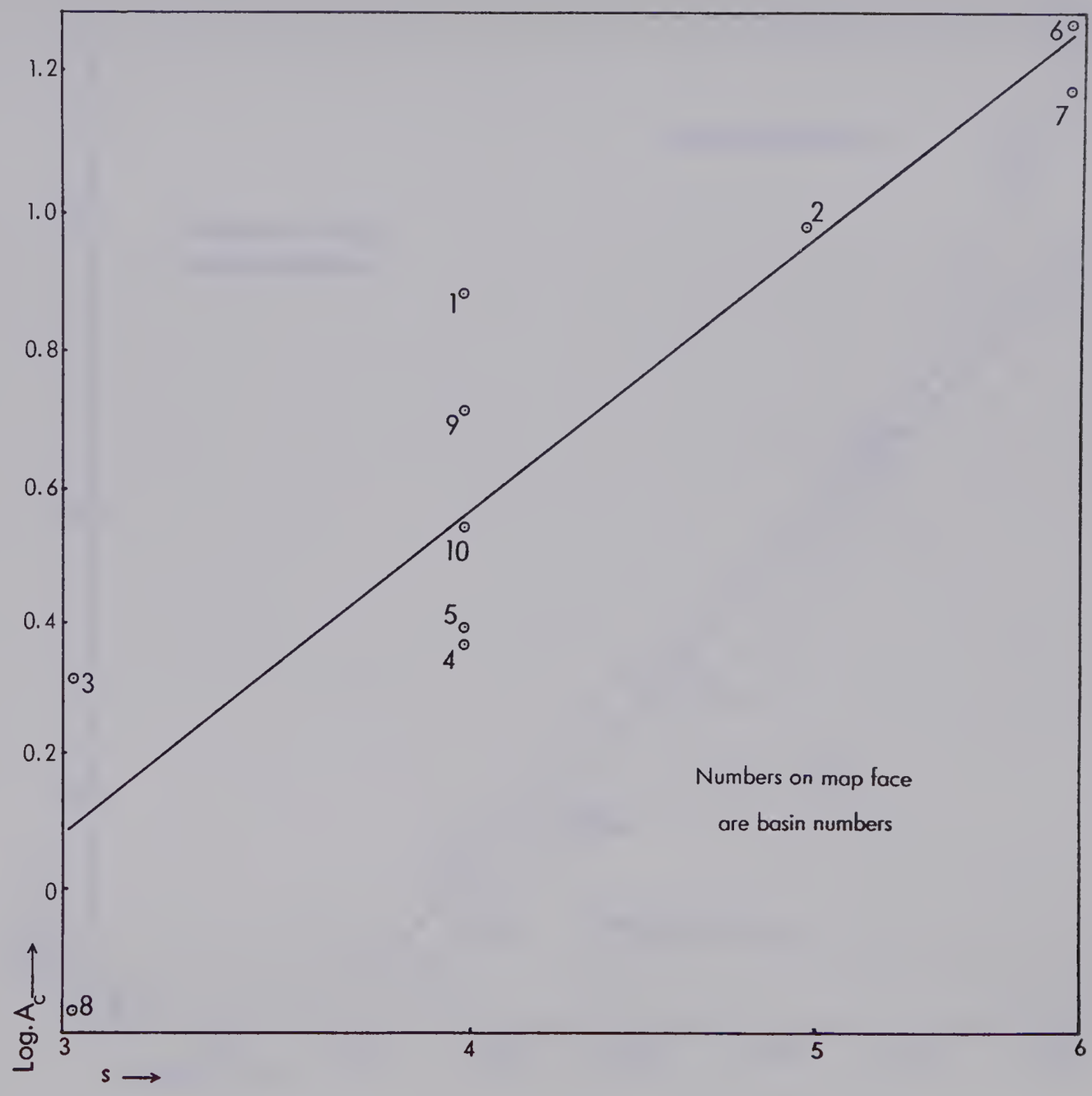
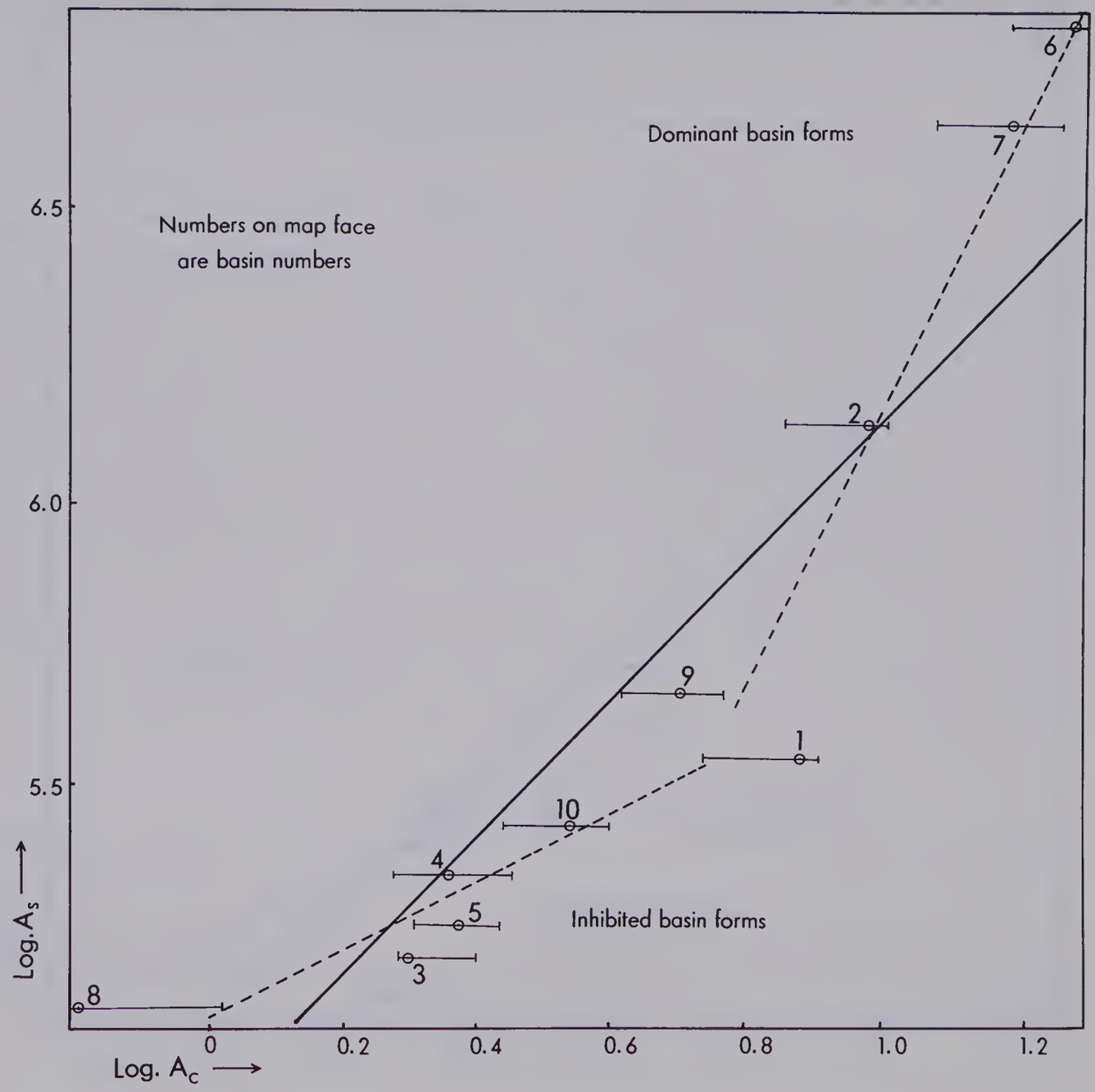




Fig. 46

Linear regression
Log. A_S with Log. A_C



Relationship between plotted variables:

$$A_s = 1.7(A_c)^{1.4}$$

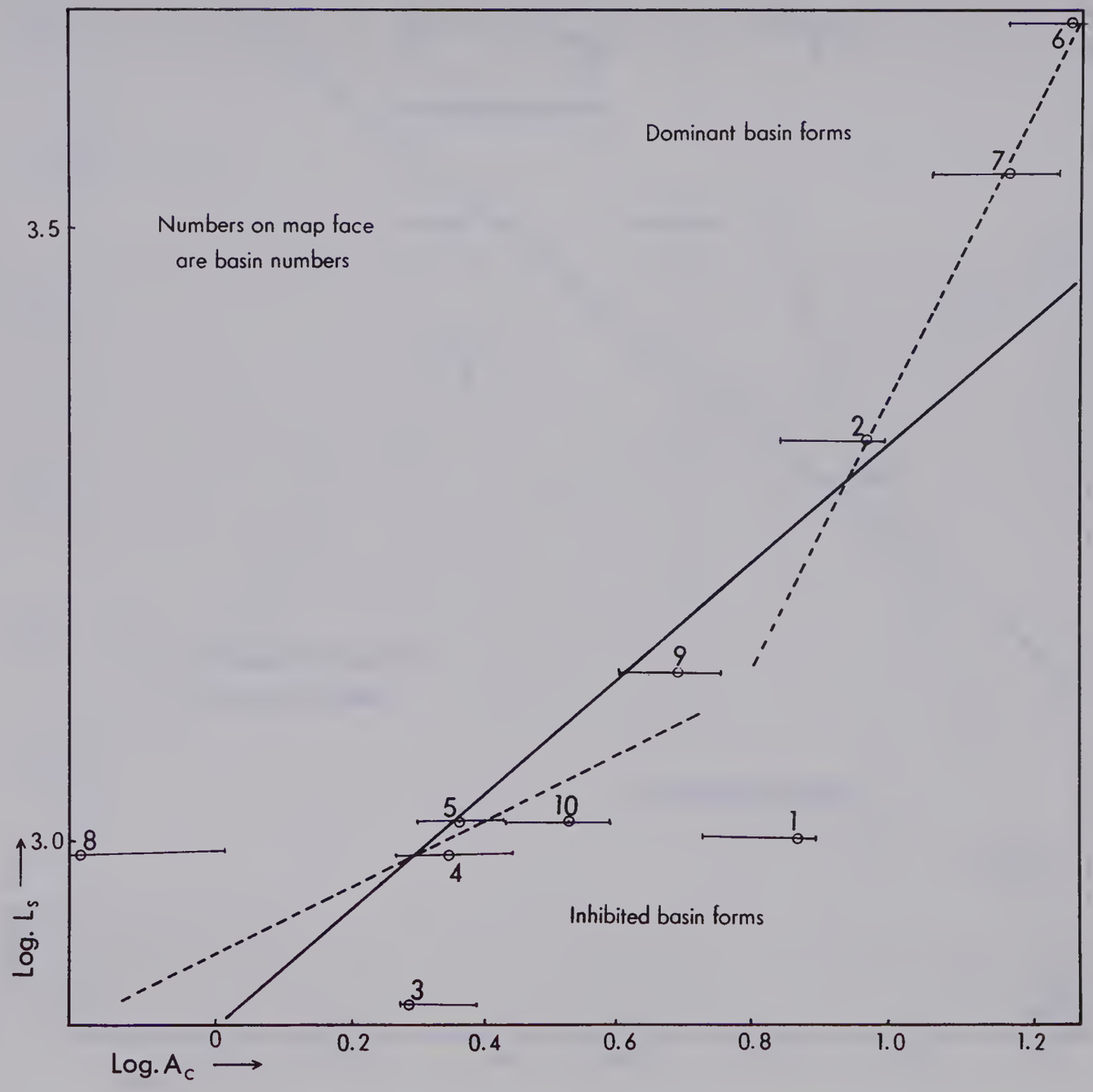
significant at 0.1% level

spread of values around the mean



Fig. 47

Linear regression
Log. L_s with Log. A_c



Relationship between plotted variables:

$$L_s = 594.6(A_c)^{0.5}$$

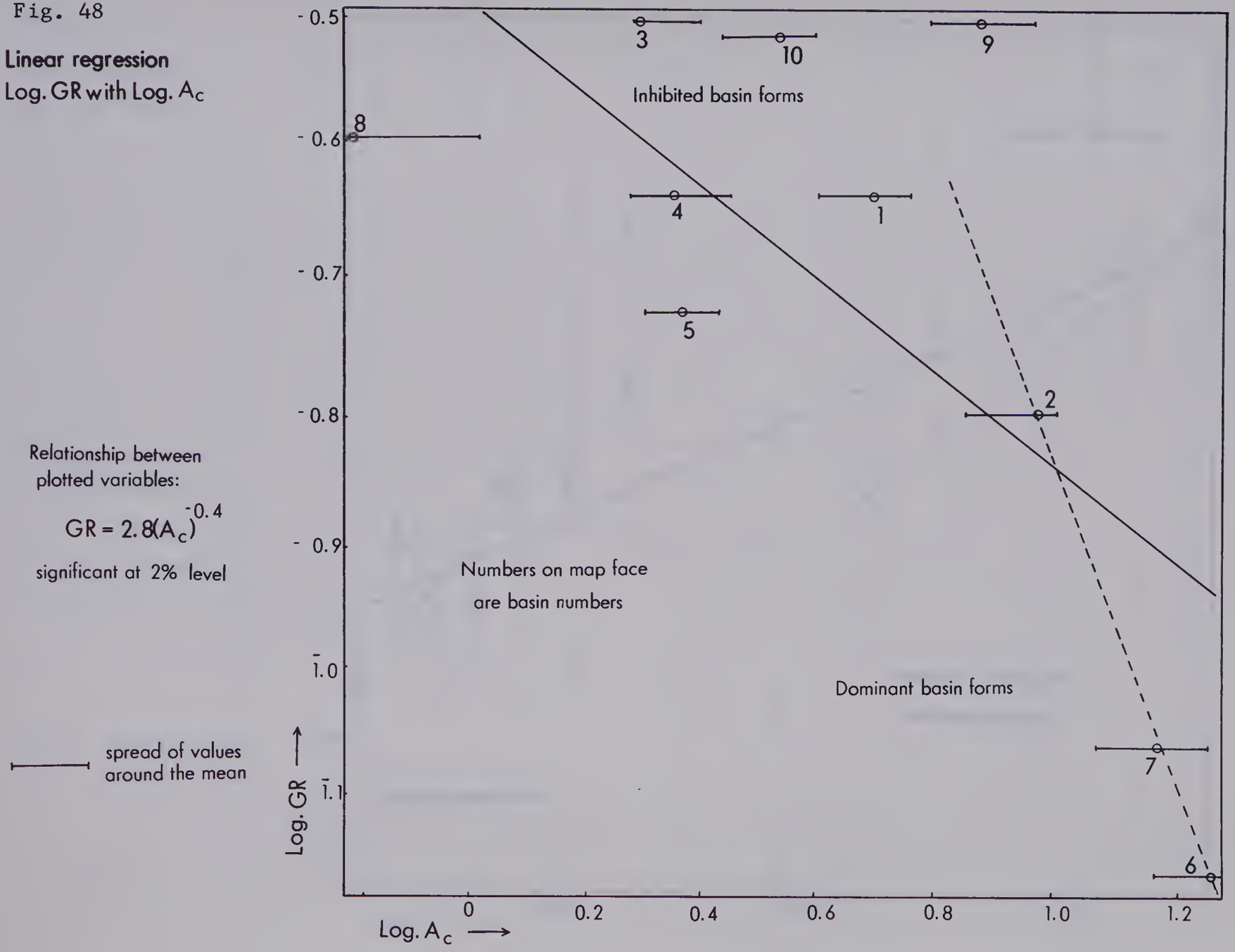
significant at 1.0% level

spread of values around the mean



Fig. 48

Linear regression
Log. GR with Log. A_c



Relationship between plotted variables:

$$GR = 2.8(A_c)^{-0.4}$$

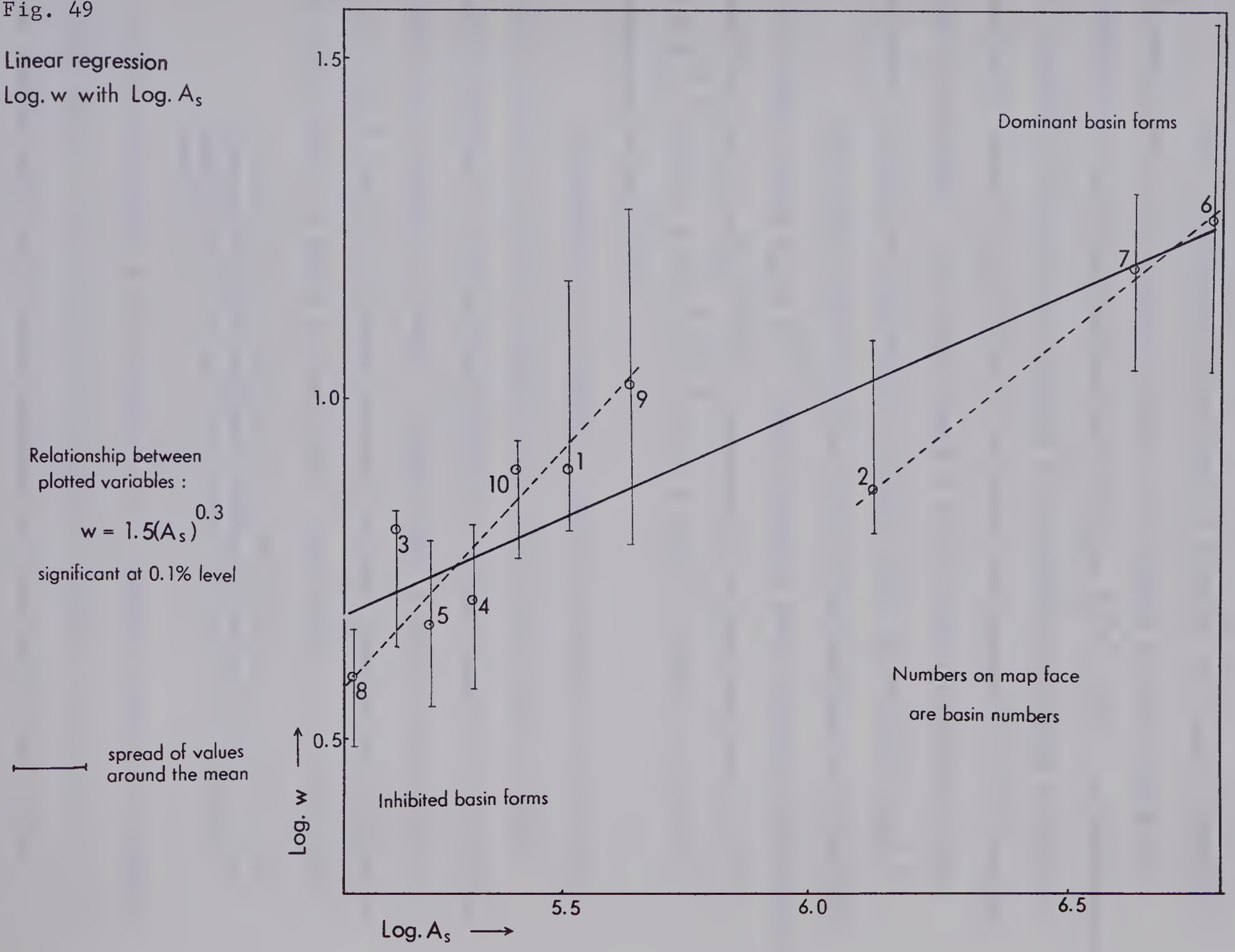
significant at 2% level

spread of values around the mean



Fig. 49

Linear regression
Log. w with Log. A_s



Relationship between plotted variables :

$$w = 1.5(A_s)^{0.3}$$

significant at 0.1% level

spread of values around the mean



general points fit well around the fitted line ($w \alpha A_s^{0.3}$), (7), which is significant at the 0.1% level.

Fig. 50 illustrates the rate of increase of computed bankfull discharge with increasing basin size. All points fit well along the regression line, whose gradient is 0.6 ($Q_{bc} \alpha A_s^{0.6}$), (8). An examination of equation (3) shows this exponent to be somewhat less than that documented by Leopold, et al (1964), but since the exponent in this equation has in other work decreased from 0.8 to 0.7 with increasingly arid climates, 0.6 can be considered a feasible value in the study area.[1]

From these regression lines other relationships can be derived. For example, since $Q_{bc} \alpha A_s^{0.6}$, then obviously the reciprocal relationship must be true, $A_s \alpha Q_{bc}^{1.6}$ (9). Combining this last equation with (7), $w \alpha A_s^{0.3}$, then $w \alpha (Q_{bc}^{1.6})^{0.3}$, or $w \alpha (Q_{bc})^{0.5}$, which is the well-documented equation (4).

In examination of (b+f+m), Leopold and Maddock (1953) noted that velocity increases in general much slower than width and depth with discharge in a downstream direction (m often equal to 0.1 or thereabouts for perennial streams). In some cases V may be constant (m = 0, $Q_b = k'[A_c]$, where k' is some constant) or even increase with discharge, in which case m is small and negative. Dury (1969, p. 326) noted that,

> "...theoretical analysis of stream behavior at bankfull combines with pilot empirical studies of velocity at $Q_{1.58}$ to suggest that constant downstream velocity is attained at bankfull."

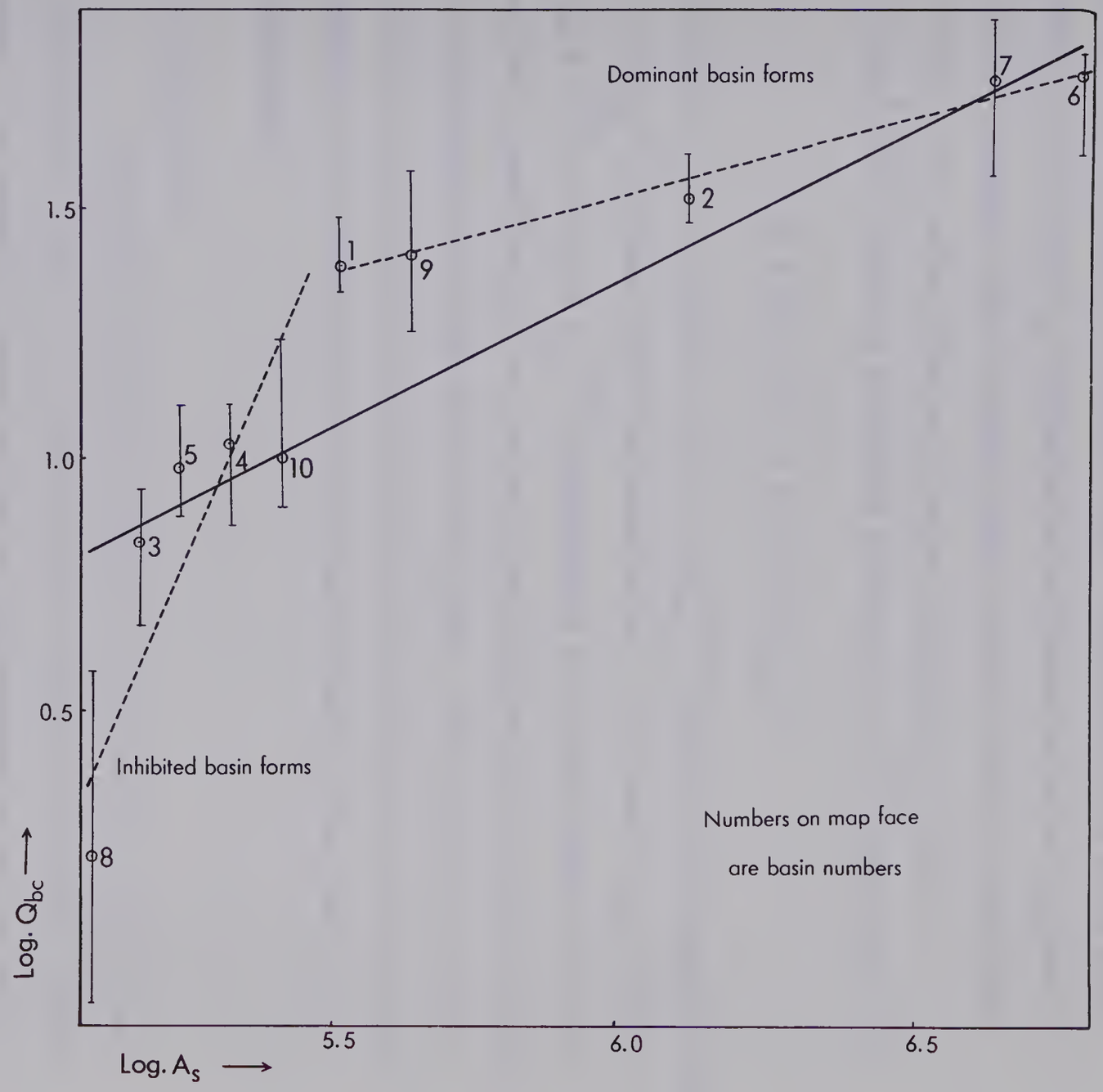
so that in this case m = 0, and b+f = 1.0. However, Leopold and Maddock (1953) found m = 0.2 for $Q_{2.3}$ on ephemeral channels in New Mexico. In the present study, all channel segments were not part of one system

[1] Pers. comm. L. B. Leopold, March 6th, 1970.

Fig. 50

Linear regression

Log. Q_{bc} with Log. A_s



Relationship between plotted variables :

$$Q_{bc} = 0.09(A_s)^{0.6}$$

significant at 0.1% level

spread of values around the mean



taken in a downstream direction, but separate basins, some of which experience form control. Gradient lowers considerably with basin size. Thus from values of V_{bc} derived using the Manning equation, V_{bc} was found to decrease with increasing area, and thus with increasing discharge. This can be illustrated in the following way. Since $A_s \alpha A_c^{1.4}$ (5), then the reciprocal is true, $A_c \alpha A_s^{0.7}$. Thus from (9),

$$A_c \alpha (Q_{bc}^{1.6})^{0.7}, \text{ i.e. } A_c \alpha Q_{bc}^{1.1}.$$

Because $w \cdot d = A_c$, and $w^b \cdot d^f \cdot V^m = Q_{bc}^{b+f+m}$ then it can be deduced that $Q_{bc} \alpha V_{bc}^{-0.1}$ because b+f+m must equal unity. As a result, $Q_{bc} \alpha w^{0.5}$, $Q_{bc} \alpha d^{0.6}$, and $Q_{bc} \alpha V_{bc}^{-0.1}$ approximately in the study area. In this way the channel geometry of the ten study segments can be totally related to the bankfull discharge for the ten streams. Thus, since Q_{bc} is also related to A_s (Fig. 50) then the inter-relationship between the basin morphometry and the development of the channels draining these basins are established.

B. Conclusions

The basins developing peripheral to the Red Deer River in the study area are doing so in closely similar climatic and geologic environments. The semiarid climate results in high evaporation and low infiltration, this latter being predominantly in the form of piping. Annual precipitation is low, from 12 inches to 16 inches, and falls mainly in the summer months. Runoff is seasonal, having a maximum in the spring from snow melt, and again in the summer as a result of high intensity, short duration storms. These storms rapidly erode the non-fissile claystones and fine sandstones, which are almost horizontally interbedded,

into typical badland scenery. Thus, the summer sediment yields are high from the rapidly developing gullies.

The high rate of erosion inhibits vegetation, and allows factors controlling the development of individual slope elements to be at least partly elucidated. A theory of slope evolution for each element as it occurs in the vertical sequence can consequently be proposed.

Despite the consistency in the individual slope evolution mechanisms across the area, the erosional form of the individual basins is not the same. This is because basins are evolving with respect to two controlling facets, the retreating scarp face from the alluvial flats to the upper prairie, and the upper prairie itself. Small basins evolving on the retreating face that have not breached the upper prairie level are inhibited by the larger basins which have, and thus the former experience greater geological control which is evident in their hypsometric curves and longitudinal stream profiles. A study of the latter two characteristics leads to model of basin evolution as basin size increases.

Within the ten basins of different size and erosional development, a surprising degree of planimetric consistency exists. This is apparent in analysis of the drainage net using the Strahler (1952, modified Horton) method, whereby the number, length, and area drained by streams of a given order are regressed with order for each of the basins. Deviations from the prescribed regressions, and a study of the morphometric ratios, can be explained within the evolutionary model, some basins being severely truncated in their middle orders due to the inhibition of their form. Average length of streams in basins of all orders correlates well with the average area drained by this length of stream, thus establishing a key connection between L_d (stream length) and A_d

(drainage area) throughout the area. More specifically, the length L_s of the main channel at each of the channel study segments located at the mouth of the basin correlates well with the drainage area above this point, A_s, particularly for the dominant basins.

The characteristics of the channel segments themselves can be mapped in the field, and the interdependence of all hydraulic variables within the channel described from a study of the channel form. Bankfull velocity and discharge can be computed using the Manning equation, and the variations of these along the segment can be seen as a direct response to observable changes in channel cross-sectional area, form ratio, sinuosity and roughness.

The mean cross-sectional area of the channel, A_c, and width, w, and the mean value of the computed bankfull discharge, Q_{bc}, can be shown to correlate well with key variables within the drainage net; mainstream length, L_s and gradient, GR, and the basin area, A_s. Some general trends within the regression lines may be explicable in terms of the suggested model of basin evolution. From the equations fitted to the regressions, other relationships are derived that further clarify the laws of proportionality as they pertain to the region. All the derived equations are closely correlated with those resulting from work done in similar environments.

The Steveville area, being an interacting system of tributary gullies incised into a plain of uniform elevation composed of laterally homogeneous lithology and controlled by a common base level, produces an almost ideal geomorphic model.

This model, and the forms and patterns examined here, provides additional evidence for the inter-relationships between channel morphology

and basin morphometry already established and reveals new information on the sequential evolution of badlands drainage systems.

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Appendix A

Basin #	O	Total # Streams	r_b	Total Length	Average Length	r_l	Average Area	r_A
	1	27		2008	72		7680	
			3.4			2.0		2.5
	2	8		1184	144		19200	
1			2.7			0.9		2.1
	3	3		384	128		40320	
			3.0			4.4		8.1
	4	1		560	60		326520	
			Mean 3.0			Mean 2.5		Mean 4.2
	1	139		11944	80		8320	
			4.0			1.8		4.0
	2	35		5016	144		33280	
			5.0			2.1		4.0
2	3	7		2080	296		134400	
			3.5			1.7		2.8
	4	2		1016	504		375640	
			2.0			1.6		3.7
	5	1		784	784		1262976	
			Mean 3.8			Mean 1.8		Mean 3.6
	1	26		2256	88		8320	
			4.3			1.2		3.7
3	2	6		648	104		30720	
			6.0			3.8		4.8
	3	1		392	392		148480	
			Mean 5.1			Mean 2.5		Mean 4.2

Basin #	O	Total # Streams	r_b	Total Length	Average Length	r_l	Average Area	r_A
	1	32		2000	64		8320	
			4.6			1.2		1.8
	2	7		600	80		15360	
4			3.5			1.1		3.1
	3	2		176	88		48000	
			2.0			5.0		4.4
	4	1		440	440		211840	
			Mean 3.3			Mean 2.4		Mean 3.1
	1	36		2912	80		7680	
			5.1			2.2		2.5
	2	7		1256	176		19200	
5			2.3			0.7		3.7
	3	3		344	122		70400	
			3.0			3.7		2.5
	4	1		456	456		174720	
			Mean 3.4			Mean 2.2		Mean 2.9
	1	466		44608	96		8307	
			3.5			1.5		2.8
	2	134		20280	144		23040	
			5.0			2.0		3.4
	3	27		7992	296		78080	
6			4.5			1.3		4.3
	4	6		2424	400		335360	
			3.0			2.5		3.8
	5	2		2008	1000		1263360	
			2.0			1.3		4.7
	6	1		1288	1288		6009600	
			Mean 3.6			Mean 1.7		Mean 3.8

Basin #	O	Total # Streams	r_b	Total Length	Average Length	r_l	Average Area	r_A
	1	354		35168	96		8320	
			3.5			1.6		2.8
	2	101		15464	152		23040	
			4.8			2.0		3.3
	3	21		6400	304		76800	
7			4.2			1.5		4.4
	4	5		2344	464		335360	
			2.5			2.1		3.8
	5	2		2008	1000		1263360	
			2.0			0.1		3.3
	6	1		80	80		4110464	
			Mean 3.4			Mean 1.6		Mean 3.5
	1	17		1104	64		6420	
			4.2			2.8		3.2
8	2	4		704	178		20480	
			4.0			2.4		5.8
	3	1		424	424		119680	
			Mean 4.1			Mean 2.6		Mean 4.5
	1	51		4056	88		8960	
			3.9			1.6		2.9
	2	13		1776	144		25600	
9			4.3			3.3		2.5
	3	3		1623	480		64000	
			3.0			1.4		6.8
	4	1		656	656		435200	
			Mean 3.7			Mean 2.1		Mean 4.0

Basin #	O	Total # Streams	r_b	Total Length	Average Length	r_l	Average Area	r_A
	1	39		2336	56		10880	
			3.9			2.0		2.6
	2	10		1120	112		28800	
10			3.3			2.1		2.9
	3	3		704	232		84480	
			3.0			2.8		3.0
	4	1		656	656		254720	
			Mean 3.4			Mean 2.3		Mean 2.8

Appendix B

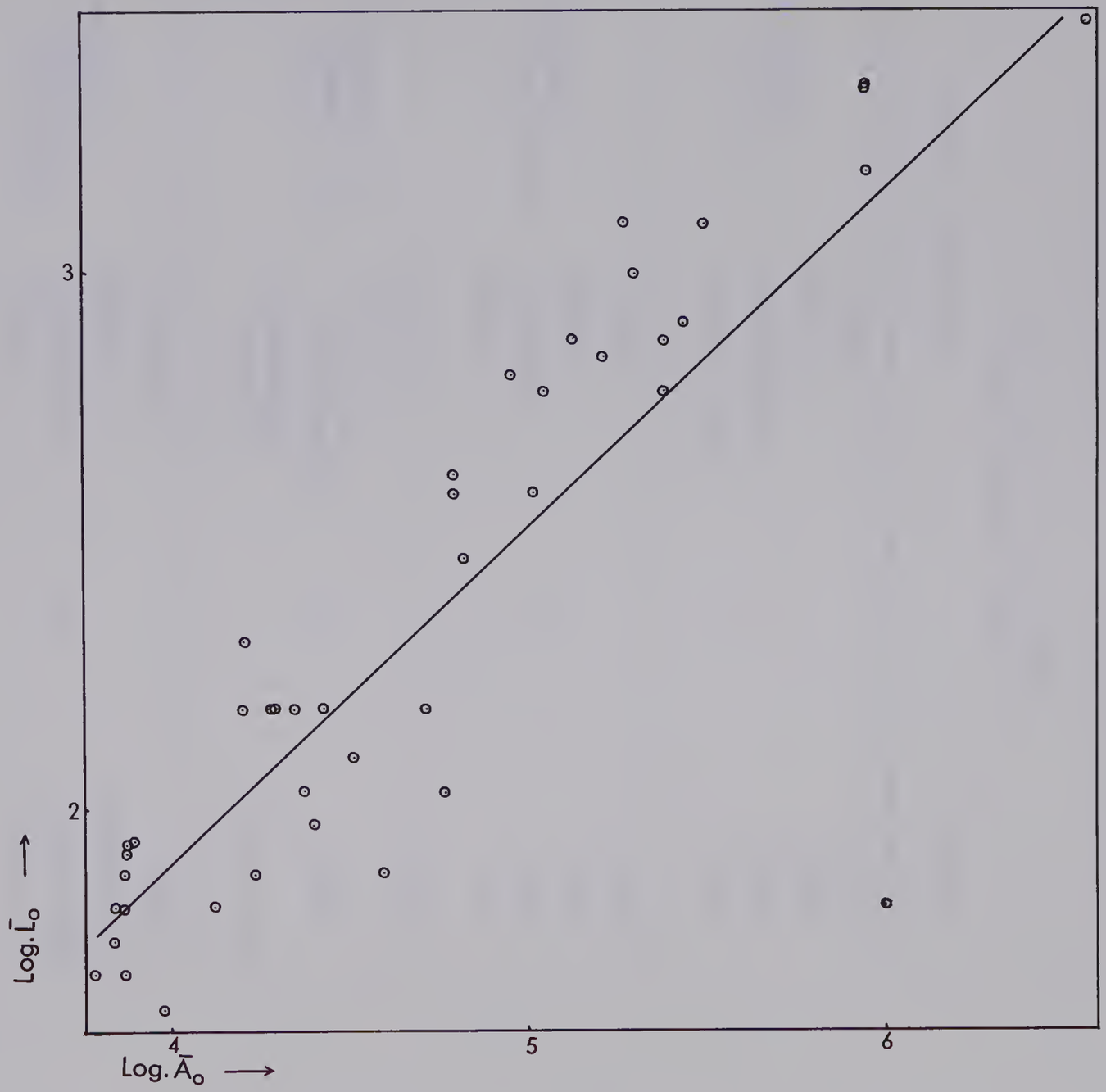
Linear regression.

Log. $\bar{L}_o$ with Log. $\bar{A}_o$

Relationship between plotted variables :

$$\bar{L}_o = 1.6(A_o)^{0.4}$$

significant at 1% level





Appendix C

Values for the Computation of the Roughness Coefficient

$$n = (n_o + n_1 + n_2 + n_3 + n_4)m_5$$

After Te Chow, (1964)

Channel Conditions			Values
Material involved	Earth	n_o	0.020
	Rock cut		0.025
	Fine gravel		0.024
	Coarse gravel		0.028
Degree of irregularity	Smooth	n_1	0.000
	Minor		0.005
	Moderate		0.010
	Severe		0.020
Variations of channel cross-section	Gradual	n_2	0.000
	Alternating occasionally		0.005
	Alternating frequently		0.010-0.015
Relative effect of obstructions	Negligible	n_3	0.000
	Minor		0.010-0.015
	Appreciable		0.020-0.030
	Severe		0.040-0.060
Vegetation	Low	n_4	0.005-0.010
	Medium		0.010-0.025
	High		0.025-0.050
	Very high		0.050-0.100
Degree of meandering	Minor	m_5	1.000
	Appreciable		1.150
	Severe		1.300

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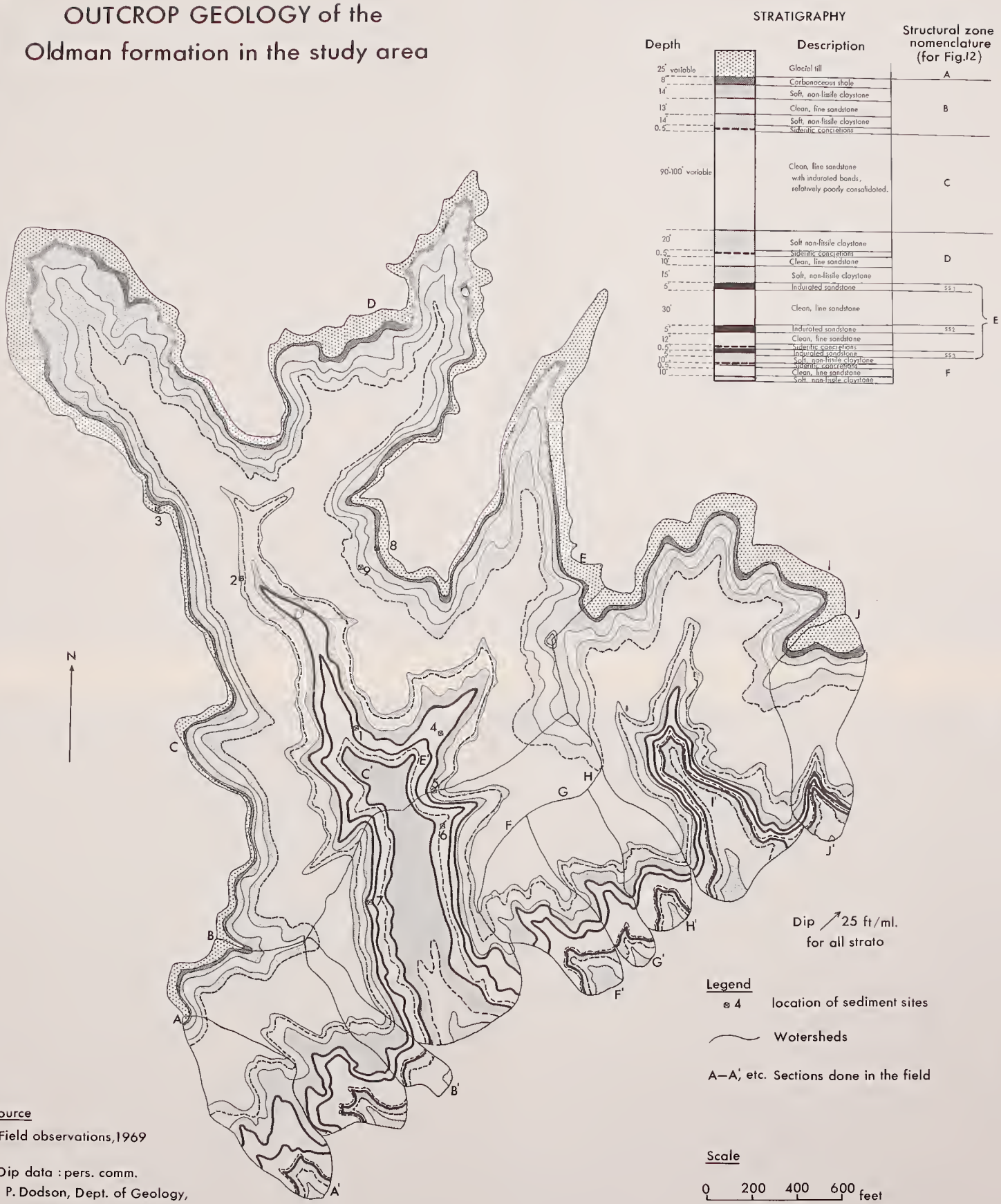
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Fig. 3

OUTCROP GEOLOGY of the Oldman formation in the study area



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Field observations, 1969

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University of Alberto.

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